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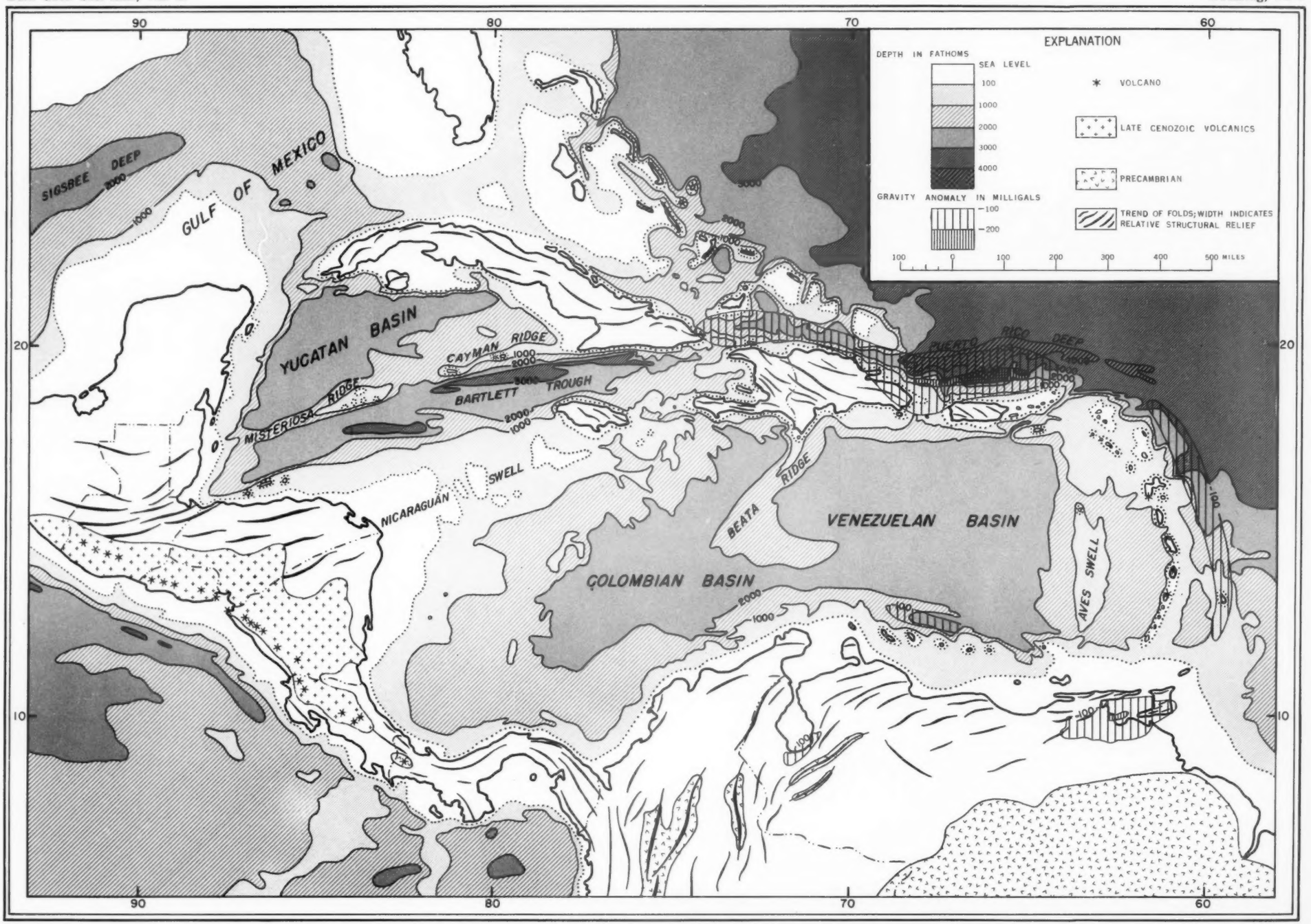

GEOGRAPHIC AND STRUCTURAL SETTING OF CARIBBEAN REGION

Base, submarine contours, and iosgam contours from de Bruyn (1951), by permission of E. J. Brill, Leiden

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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CARIBBEAN LAND AND SEA THROUGH THE AGES

(Address of the Retiring President of The Geological Society of America)

BY W. P. WOODRING

ABSTRACT

The oldest part of the Caribbean region proper is in northern Central America, where Permian(?) and Lower Permian marine deposits rest on metamorphic rocks of unknown, possibly middle Paleozoic, age. According to present dating, geosynclinal deposition spread eastward in Late Jurassic time to include Cuba, farther eastward and southward in Early Cretaceous time to include Hispaniola and probably Jamaica, and still farther eastward in Late Cretaceous time to include Puerto Rico, the Virgin Islands, and St. Croix.

Throughout the Caribbean region, the Cretaceous is characterized by volcanics of great thickness, pyroclastics being more widespread and thicker than flows. These volcanics evidently were derived from lands of unknown size that are now under the waters of the Caribbean Sea. Land still persisted south of eastern Cuba during Eocene time. Thereafter no geological evidence is now available pointing to land in the Caribbean Sea.

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INTRODUCTION

Paleogeography starts as a concoction of essential ingredients, generally too meager, and winds up as a heady essence distilled through the imagination of the perpetrator. It therefore is an art. It is practiced—and too frequently abused—by geologists, paleontologists, and biogeographers. It is a simple matter to make paleogeographic generalizations, but it is something else to lay them down on a map, even on a small-scale map, and practically all paleogeographic maps are of very small scale. On a map your ideas look up at you coldly, knowing full well they will come back to taunt you.

For the most part published paleogeographic maps of the Caribbean region are based on the premise that almost the entire Caribbean Sea has indeed been sea since the earliest dated events in the geological history of the region. Schuchert's (1935) well-known and widely used maps, for example, show land of varying width extending from Central America northeastward to Jamaica and Cuba, but the re-

mainder of the Caribbean Sea is enduring sea. That premise has been challenged during recent years (Bucher, 1947, p. 100; Eardley, 1952). It is not my intention to present a long series of paleogeographic maps: that would be dull for any audience. I propose to present a few maps representing intervals of time selected to test the validity of the premise on the basis of what is now known about the geology of the Caribbean region.

Almost any conclusions, at least concerning the pre-Tertiary history, are bound to be tentative. Our knowledge of the pre-Tertiary geology of the Greater Antilles is meager. There has been plenty of reconnaissance, and generalized small-scale maps are available, but very little real mapping has been published. In all of the Greater Antilles, an area of 80,000 square miles, geological maps of pre-Tertiary areas on a scale of a mile to the inch or better are available for about 600 square miles (Flint, de Albear, and Guild, 1948, Pl. 18; Matley, 1951). To be sure, the pre-Tertiary rocks of Puerto Rico are shown on maps having a scale of 1:62,500 or 1:50,000, published during the period 1919–1931. Despite the scale, they are reconnaissance maps. Though ultramafic rocks crop out in all the provinces of Cuba, one of the maps just cited (Flint, de Albear, and Guild, 1948, Pl. 18) is the only published detailed map showing any part of them, and it covers a minute segment. Another map (Koschmann and Gordon, 1950 [1951], Pl. 19), covering 350 square miles, is the only map on any scale showing a considerable number of pre-Tertiary units in the entire island of Hispaniola. It represents reconnaissance mapping on a scale of 1:100,000. It is based, however, on mapping and not on generalizations from hasty road and trail traverses.

Though the Caribbean region includes Central America and northern South America, this account deals principally with the Antilles.

Geographic and Structural Setting of Caribbean Region

About 90 per cent of the Caribbean region, exclusive of the continental mainland, is under water. All that is known about that part is its general configuration; the results of gravity observations at about 100 stations; and the results of geophysical probing of the thickness of sediments in one area: the Puerto Rico Deep.

Plate 1 shows the gross submarine features. The greatest depths are in the Puerto Rico Deep (4350 fathoms) and in the Bartlett Trough (just short of 4000 fathoms), directly south of the highest range in Cuba (the Sierra Maestra) and directly south of the highest peak in that range.

Plate 1 also shows areas of pronounced negative gravity anomalies—anomalies greater than 100 milligals. The isogams of Plate 1 are based on de Bruyn's (1951) 1:10,000,000 25-isogam map, the most recent compilation of gravity data covering the Caribbean region. The Vening Meinesz narrow belt of strong negative anomalies extends from eastern Cuba to the eastern Venezuelan basin. The largest anomalies are just south of the bottom of the Puerto Rico Deep (−219 milligals) and in the eastern Venezuelan basin (a little more than −200 milligals). According to de Bruyn's map, the narrow belt of negative anomalies north of the Venezuelan coast, shown by an incomplete −100 isogam on Plate 1, evidently is not continuous with the Vening Meinesz belt. Though data are still lacking, the belt off the Venezuelan coast may bend northeastward and flare out up the Aves Swell.

The Precambrian areas of Plate 1 are at the far south border of the Caribbean region. The metamorphic rocks of the Guianan Shield, the counterpart of the Canadian Shield, in the extreme southeastern part of the map, are overlain by lower Paleozoic deposits 600 miles to the southeast in Brazil. The smaller of the two areas in the Venezuelan Andes evidently is older than the Middle Ordovician near by (Leith, 1938, p. 337–339) and therefore is presumably Precambrian. The larger area is assumed to be of the same age, but the oldest fossiliferous strata overlying the metamorphic rocks are of Devonian age. The metamorphic rocks of the Cordillera Central of the Andes in Colombia are overlain by Middle Cambrian to Middle Ordovician 125 miles south of the south border of Plate 1 (Harrington and Kay, 1951, p. 655-657 and by Lower Ordovician within the area of Plate 1 (Botero Arango, 1942, p. 24). The metamorphics of the Cordillera Oriental may possibly be Paleozoic. The oldest dated deposits

overlying them are nonmarine strata of early Mesozoic age.

The late Cenozoic volcanics of Central America (Pl. 1) form a distinct geological province. They are likely to remain the chief obstacle in attempts to integrate the geology of southern Central America with that of northern Central America.

The trend and relative structural relief of folds are represented on Plate 1. Faults, however, are omitted. Many faults mapped on land could not be adequately shown on the scale of that illustration. Some of the submarine faults that have been suggested need confirmation.

Outline of Geological History of Caribbean Region

Pre-Jurassic

The oldest part of the Caribbean region proper is not in the Antilles; it is in the folded mountain ranges of northern Central America. Gneiss, schist, and other metamorphic rocks, exposed in the cores of anticlines in central Guatemala, northern Honduras, and northern Nicaragua, are the oldest rocks in northern Central America. They have been assigned to the Precambrian since the pioneer work in that area. All that is now known about their age is that they are older than the 2000 feet or more of Permian(?) and Lower Permian marine strata resting on them. These ancient-appearing rocks may indeed be of Precambrian or early Paleozoic age. In view, however, of the geological history of the entire region, it is more probable that they are not older than middle Paleozoic. They have been through two periods of deformation and batholitic intrusion and therefore are strongly metamorphosed. The Permian(?) and Lower Permian rocks themselves appear to be progressively metamorphosed eastward: shale to phyllite to schist (Roberts and Irving, In preparation).

In the entire Caribbean region and its borders marine Triassic deposits are lacking. Marine Upper Triassic, interbedded with red beds, has been found in the eastern foothills of the Cordillera Central of the Andes in Colombia, 75 miles southwest of Bogotá (Trumpy, 1943, p. 1297–1299), but not farther north. The widespread nonmarine strata of Colombia and Venezuela, which overlie marine upper Paleozoic and underlie marine Lower Cretaceous, are thought to represent much of Triassic and Jurassic time.

Sometime during the interval between Middle Permian and Late Triassic, the pre-Permian and Permian strata of northern Central America were deformed and intruded by ultramafic and granitic rocks (Roberts and Irving).

Jurassic

After the deformation in northern Central America just mentioned, nonmarine deposits were laid down in that area north and south of the folded ranges, which trend east-northeastward in an arc of gentle curvature, and in intermontane basins. These nonmarine deposits contain plants of Early and Middle Jurassic age. In Guatemala and the adjoining Mexican state of Chiapas, the upper part of the nonmarine strata include gypsum and salt (Sapper, 1937, p. 27), suggesting that marine Middle or Upper Jurassic underlies northern Guatemala and Yucatán. There is a still undocumented report of marine Upper Jurassic in Honduras (Imlay, 1952, p. 969–970).

Marine Lower and Middle Jurassic, like marine Triassic, is unknown in the Caribbean region proper. Marine Lower Jurassic has been found at a locality along the south border of the region in Colombia: near the north end of the Cordillera Central of the Andes (Trumpy, 1943, p. 1299).

By middle Late Jurassic time (late Oxfordian; Imlay, 1952, p. 969) geosynclinial deposition reached the Greater Antilles, but only the westernmost island, Cuba. Thick marine deposits of that age, including rocks mildly metamorphosed to phyllite and quartzite, are found in westernmost Cuba. The much more strongly metamorphosed rocks of Isla de los Pinos (L. M. R. Rutten, 1934; Page and McAllister, 1944, p. 181–184), off the south coast of western Cuba, and the similar pre-ultramafic rocks of central and eastern Cuba, may possibly be of the same age. If they are, deformation and intrusion of diorite, now more or less gneissic, took place during a span from early Kimmeridgian to early Portlandian time, corresponding to a gap in the

Cuban sequence (Imlay, 1942, p. 1439–1440). Late Portlandian (Imlay, 1942, p. 1433–1440; 1952, p. 969) is represented in western, central, and eastern Cuba by 2000 feet of limestone, shale, chert, and some tuff.

Early Cretaceous

Early Cretaceous geosynclinal deposition took in all of Cuba and Hispaniola, probably Jamaica, and possibly Puerto Rico. If it included Jamaica, the deposits are represented in the metamorphic rocks of the Blue Mountains. This is the beginning of enormously thick volcanics: mostly pyroclastics in Cuba, mostly flows (including submarine flows) in Haiti, mostly pyroclastics in the Dominican Republic.

All the major subdivisions of the Lower Cretaceous are reported to be represented in Cuba (MacGillavry, 1937, p. 7–13; Maync, 1949, p. 531–532), but no records of a well-dated sequence have so far been published. Both the late Portlandian and the earliest Lower Cretaceous (Neocomian) of Cuba—the Aptychi limestones or Aptychi formation of Dutch geologists (MacGillavry, 1937, p. 7–8; M. G. Rutten, 1936, p. 10–12; van Wessem, 1943, p. 6–8)—contain concentrations of ammonite opercula, radiolaria, and pelagic foraminifera. They evidently are fairly deep-water deposits. The presence of land-plant debris is not opposed to that interpretation. Great quantities of decomposed land-plant debris were dredged by the ALBATROSS in deep water off the Pacific coast of Central America (Agassiz, 1892, p. 11–12). The Cuban deposits are typically geosynclinal, typically eugeosynclinal, if that term is preferred.

The Lower Cretaceous of most of Cuba is made up principally of marine pyroclastics, with which minor flows are interbedded: part of the Tuff formation or Tuff series of Dutch geologists (de Vletter, 1946, p. 11–15; Hermes, 1945, p. 3–5; MacGillavry, 1937, p. 8–10; M. G. Rutten, 1936, p. 7–10; Thiadens, 1937, p. 11–18; van Wessem, 1943, p. 8–12; Vermunt, 1937, p. 15–17). Along the north coast of the island, in the northwestern part of Camagüay Province, van Wessem's (1943, p. 7–8) Aptychi formation includes thick beds of gypsum. Northward from the north coast of Cuba, in the subsurface strata of the Bahamas and peninsular Florida, the Lower Cretaceous consists of carbonate rocks. The only deep well so far drilled in the Bahamas penetrated almost 4000 feet of presumably Lower Cretaceous carbonate deposits and did not reach their base. Some of the dolomite of the carbonate deposits contains blebs of anyhydrite. The subsurface Lower Cretaceous carbonate rocks of peninsular Florida include minor amounts of bedded gypsum or anhydrite (P. L. and E. R. Applin, 1944, p. 1721).

The ultramafic rocks of Cuba probably were emplaced at an early date in Early Cretaceous time; that is, they probably are younger than the Central American ultramafics. In the east-central part of the island, where the ultramafics have been mapped (Flint, de Albear, and Guild, 1948, p. 43–45), they engulf schist and gneiss of unknown age and are overlain by volcanics that appear to range in age from Early to Late Cretaceous, but the lower, presumably Early Cretaceous, part is not yet dated. The late Upper Cretaceous deposits contain serpentine debris (L. M. R. Rutten, 1940, p. 542).

Much of what has been assigned to the Lower Cretaceous in Haiti, with varying degrees of qualification (Woodring, Brown, and Burbank, 1924, p. 86–93), is now known to be Upper Cretaceous. Nevertheless some of the lavas, particularly the earlier basalt of the northern part of the country and the basalt of the southern peninsula, may be Lower Cretaceous. Limestone between pillows of basalt, in the extensive area of basalt southwest of Port-au-Prince, shows sections of rudists comparable to those from Lower Cretaceous limestone in the Dominican Republic.

The metamorphic rocks of the Cordillera Central of the Dominican Republic have been claimed to be pre-Cretaceous, possibly Paleozoic (Weyl, 1940, p. 5, 28–29; and other publications by the same author). In the only area that has been mapped, however, a great thickness of sericite schist, altered tuff, and greenstone—all estimated to represent a thickness of 20,000 feet, if there is no duplication—overlies more or less altered limestone containing rudists and other fossils of Cretaceous age, probably late Early Cretaceous (Koschmann and Gordon, 1950 [1951], p. 317–325). Ultramafic rocks form a sill-like mass, between altered tuff and green-

stone in the upper part of the sequence. Limestone 70 miles southeast of the mapped area contains the rudists found in the mapped area and also *Orbitolina concava texana*, which occurs in the Aptian-Albian of Cuba, northern Central America, Colombia, and Venezuela.

Matley (1929, p. 465; 1951, p. 22) was convinced that the metamorphic rocks of the Blue Mountains in eastern Jamaica are pre-Cretaceous. They are now thought, however, to be Cretaceous or possibly Jurassic (Zans, 1953, p. 2). If they are Cretaceous, they are likely to be Lower Cretaceous.

Late Cretaceous

Late Cretaceous geosynclinal deposition spread still farther eastward to include all of the Greater Antilles, the Virgin Islands, and St. Croix. Figure 1 is a representation of Late Cretaceous paleogeography. Like most paleogeographic maps, it takes in too great a time span and therefore is composite.

Throughout the Greater Antilles, the Virgin Islands, and St. Croix the Upper Cretaceous deposits include basaltic and andesitic pyroclastics and flows of great thickness. They are everywhere strongly deformed and practically everywhere are intruded by batholiths of quartz diorite or granodiorite, near some of which they are altered to schist. It was on St. Thomas, one of the Virgin Islands, where the Swedish naturalist Cleve (1871, p. 5) was able to show almost 85 years ago that the metamorphic rocks are altered Cretaceous.

The latest dated Upper Cretaceous of the Greater Antilles includes, or consists mostly of, detrital deposits of late Campanian and Maestrichtian age (Imlay, 1944, p. 1013–1015). They include thin beds of limestone containing a distinctive fauna, characterized by the remarkable rudist *Barrettia*, which has been found in Cuba, Jamaica, Haiti, Puerto Rico, and northern Guatemala.

Deformation and intrusion of quartz diorite or granodiorite took place in the Greater Antilles and the Virgin Islands toward the close of Cretaceous time. The deformation was the first of several episodes of varying intensity during the interval from late Cretaceous to Eocene. In Cuba quartz diorite intrudes volcanics of early Late Cretaceous age, apparently Cenomanian to Coniacian (part of the Tuff formation or Tuff series of Dutch geologists; Imlay, 1944, p. 1011–1012) and contributed debris to the Campanian-Maestrichtian detrital deposits. The deformation and intrusion therefore is middle Late Cretaceous, approximately Santonian (Imlay, 1944, p. 1012). This deformation produced the arc of islands shown on Figure 1 where Cuba now is located. Discordance between the volcanics and the detrital deposits, however, is by no means uniform. At some localities there is no appreciable discordance, but the volcanics, the quartz diorite, and the ultramafic rocks were shedding debris into the late Campanian-Maestrichtian sea.

In Jamaica granodiorite intrudes the still undated metamorphic rocks of the Blue Mountains. Pebbles of granodiorite, however, are found in conglomerates of early Eocene age, but not in conglomerates interbedded with Upper Cretaceous limestone (Zans, 1953, p. 3). In Hispaniola, Puerto Rico, and the Virgin Islands deformation and plutonic intrusion took place at about the end of the Cretaceous, but the dating is not within narrow limits.

Altered basaltic and andesitic flows and tuffs, intruded by diorite, form the pre-Tertiary basement in Panamá (Woodring and Thompson, 1949, p. 227). The tuffs, like those in the Antilles, contain radiolaria and are altered to schist near some of the stocks of diorite. These volcanic rocks are shown as Tertiary volcanics on the current edition of the geological map of North America. At least part of the volcanics is dated as Late Cretaceous, on the basis of formanifera found in siliceous limestone in northwestern Panamá near the Costa Rican border.

Though the little Dutch islands off the coast of Venezuela (Aruba, Curaçao, Bonaire) are separated from the Greater Antilles by the width of the Caribbean Sea (a distance of 400 miles), their geology is Antillean. Their basement consists of a thick pile of volcanics overlain by detrital Upper Cretaceous (Molengraaff, 1929, p. 15–25; Pijpers, 1933, p. 7–38; Westermann, 1932, p. 11–37). Tuffs and cherts in the volcanic sequence contain radiolaria, as in the Greater Antilles. The three islands show

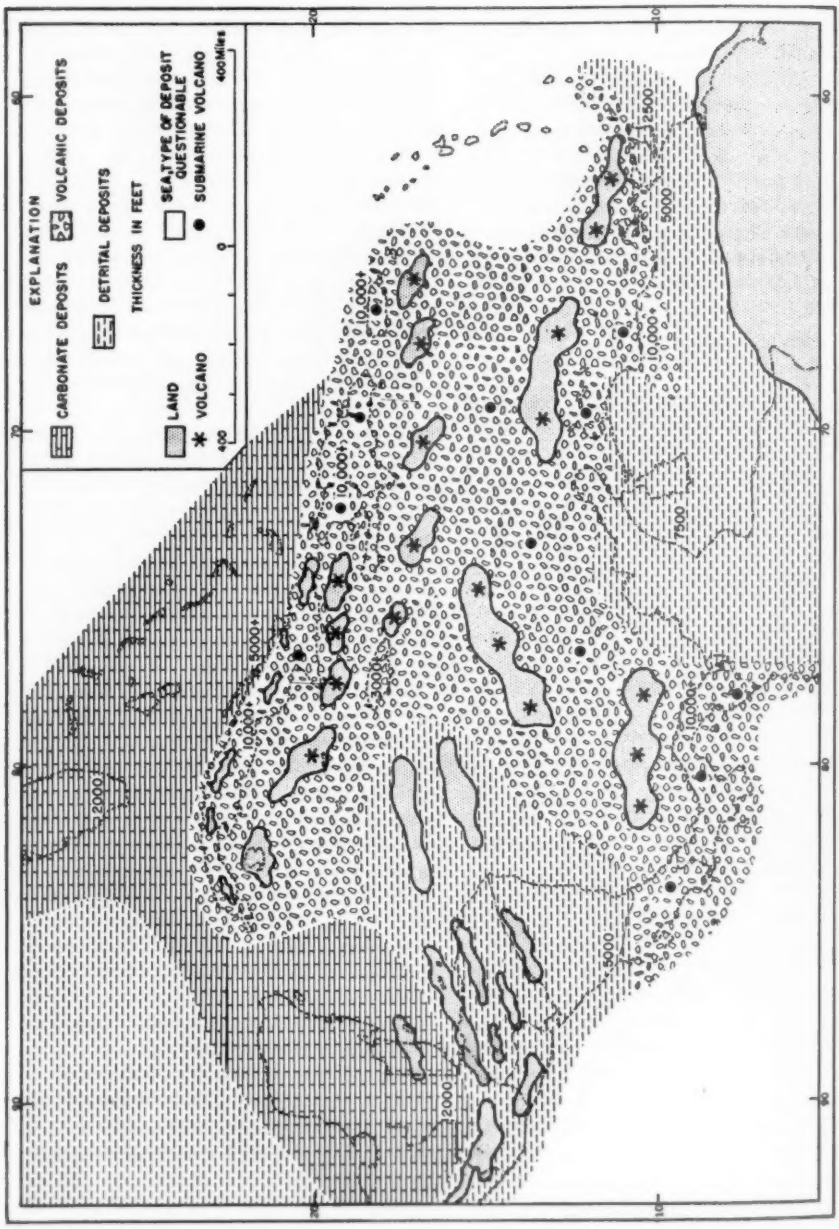

FIGURE 1.—CARIBBEAN REGION DURING LATE CRETACEOUS TIME

a westward increase in intensity of deformation, in the size of exposed stocks of quartz diorite, and in degree of metamorphism.

The highly metamorphosed rocks of the Cordillera de la Costa in northern Venezuela (Bucher, 1952, p. 39–62, 73–84) include altered volcanics and apparently represent a transition from the volcanic sequence of the Dutch islands to the nonvolcanic sequence farther south and southwest in Venezuela. Late Cretaceous (Turonian) foraminifera have been found in limestone interbedded with schist, and the entire sequence of metamorphic rocks is now thought to be of Cretaceous age, or possibly both Cretaceous and Late Jurassic. Both Late Jurassic and Cretaceous fossils have been found in the metamorphic rocks of the Northern Range of Trinidad (Sutter, 1951–52, p. 189–191, 1951), the eastern continuation of the Cordillera de la Costa.

Where was the source of this great volume of Cretaceous volcanics along the north, west, and south borders of the Caribbean Sea? Though data are too meager for estimates, pyroclastics evidently are thicker and more widespread than flows. There doubtless were some minor volcanic centers in the present Greater Antilles, but they apparently were submarine. A northern source for the pyroclastics is at once eliminated. The entire Cretaceous sequence in the Bahamas and peninsular Florida, revealed by deep wells, consists of carbonate deposits. In north-central Cuba, the widespread Upper Cretaceous volcanic facies is replaced northward by a carbonate facies (de Albear, 1947, abstract; M. G. Rutten, 1936, p. 21–24). The only source for the pyroclastics is not in sight, it is under the waters of the present Caribbean Sea. Where the volcano-bearing lands were located, what their trend was, and whether they were large or small are matters that cannot be decided on purely geological grounds and may remain speculative for an indefinite time. The representation of the volcano-bearing lands on Figure 1 is one of several possible intrpretations.

FIGURE 1.—CARIBBEAN REGION DURING LATE CRETACEOUS TIME

Paleocene and early Eocene

Paleocene deposits are mostly thin and of limited extent. In western Cuba the Paleocene is detrital (Bermúdez, 1950, p. 218–225), in southeastern Cuba volcanic (Lewis and Straczek, In preparation), in Haiti carbonate (Butterlin, 1953, p. 285). No Paleocene has been recognized so far in Jamaica or Puerto Rico.

Lower Eocene deposits, like the Paleocene, are mostly thin and not widely distributed. The lower Eocene of western and central Cuba consists of detrital and carbonate deposits (Bermúdez, 1950, p. 227–238), but in southeastern Cuba the lower Eocene is made up of volcanics, with which some limestone is interbedded (Bermúdez, 1950, p. 238–239; Lewis and Straczek). Carbonate deposits are the prevailing type in Hispaniola (Bermúdez, 1949, p. 19, 32; Butterlin, 1953, p. 285). The lower Eocene of Jamaica consists chiefly of detrital marginal marine-nonmarine deposits, and includes gypsum (Zans, 1953, p. 3). Land was near by, and the distribution of coarse detrital material suggests a source to the northeast (Trechmann, 1924, p. 4).

Strong deformation, stronger than the middle Late Cretaceous deformation, took place in western Cuba in early Eocene time (Palmer, 1934, p. 132, 141); none whatever in southeastern Cuba (Lewis and Straczek). In eastern Jamaica lower Eocene deposits are involved in intense folding (Trechmann, 1924, p. 5; 1929, p. 484; Zans, 1953, p. 3).

Middle and late Eocene

Carbonate deposits of middle and late Eocene age are widespread in the Greater Antilles and northward (Fig. 2). Both middle and late Eocene are strongly transgressive, particularly late Eocene. The carbonate deposits are thin in Cuba (Bermúdez, 1950, p. 239–263), moderately thick in Jamaica (Zans, 1953, p. 4), very thick in Hispaniola (Bermúdez, 1949, p. 10–11, 19–21, 32; Butterlin, 1953, p. 285; Woodring, Brown, and Burbank, 1924, p. 99–145[1]). If represented in Puerto Rico, they are overlapped.

The volcanic islands of the Late Cretaceous (Fig. 1) have disappeared, with the exception of remnants south of Cuba, where the deep

[1] The upper Eocene of this publication has been found, during recent years, to include middle and even lower Eocene.

mainder of the Caribbean Sea is enduring sea. That premise has been challenged during recent years (Bucher, 1947, p. 100; Eardley, 1952). It is not my intention to present a long series of paleogeographic maps: that would be dull for any audience. I propose to present a few maps representing intervals of time selected to test the validity of the premise on the basis of what is now known about the geology of the Caribbean region.

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The Precambrian areas of Plate 1 are at the far south border of the Caribbean region. The metamorphic rocks of the Guianan Shield, the counterpart of the Canadian Shield, in the extreme southeastern part of the map, are overlain by lower Paleozoic deposits 600 miles to the southeast in Brazil. The smaller of the two areas in the Venezuelan Andes evidently is older than the Middle Ordovician near by (Leith, 1938, p. 337–339) and therefore is presumably Precambrian. The larger area is assumed to be of the same age, but the oldest fossiliferous strata overlying the metamorphic rocks are of Devonian age. The metamorphic rocks of the Cordillera Central of the Andes in Colombia are overlain by Middle Cambrian to Middle Ordovician 125 miles south of the south border of Plate 1 (Harrington and Kay, 1951, p. 655–657 and by Lower Ordovician within the area of Plate 1 (Botero Arango, 1942, p. 24). The metamorphics of the Cordillera Oriental may possibly be Paleozoic. The oldest dated deposits

overlying them are nonmarine strata of early Mesozoic age.

The late Cenozoic volcanics of Central America (Pl. 1) form a distinct geological province. They are likely to remain the chief obstacle in attempts to integrate the geology of southern Central America with that of northern Central America.

The trend and relative structural relief of folds are represented on Plate 1. Faults, however, are omitted. Many faults mapped on land could not be adequately shown on the scale of that illustration. Some of the submarine faults that have been suggested need confirmation.

Outline of Geological History of Caribbean Region

Pre-Jurassic

The oldest part of the Caribbean region proper is not in the Antilles; it is in the folded mountain ranges of northern Central America. Gneiss, schist, and other metamorphic rocks, exposed in the cores of anticlines in central Guatemala, northern Honduras, and northern Nicaragua, are the oldest rocks in northern Central America. They have been assigned to the Precambrian since the pioneer work in that area. All that is now known about their age is that they are older than the 2000 feet or more of Permian(?) and Lower Permian marine strata resting on them. These ancient-appearing rocks may indeed be of Precambrian or early Paleozoic age. In view, however, of the geological history of the entire region, it is more probable that they are not older than middle Paleozoic. They have been through two periods of deformation and batholitic intrusion and therefore are strongly metamorphosed. The Permian(?) and Lower Permian rocks themselves appear to be progressively metamorphosed eastward: shale to phyllite to schist (Roberts and Irving, In preparation).

In the entire Caribbean region and its borders marine Triassic deposits are lacking. Marine Upper Triassic, interbedded with red beds, has been found in the eastern foothills of the Cordillera Central of the Andes in Colombia, 75 miles southwest of Bogotá (Trumpy, 1943, p. 1297–1299), but not farther north. The widespread nonmarine strata of Colombia and Venezuela, which overlie marine upper Paleozoic and underlie marine Lower Cretaceous, are thought to represent much of Triassic and Jurassic time.

Sometime during the interval between Middle Permian and Late Triassic, the pre-Permian and Permian strata of northern Central America were deformed and intruded by ultramafic and granitic rocks (Roberts and Irving).

Jurassic

After the deformation in northern Central America just mentioned, nonmarine deposits were laid down in that area north and south of the folded ranges, which trend east-northeastward in an arc of gentle curvature, and in intermontane basins. These nonmarine deposits contain plants of Early and Middle Jurassic age. In Guatemala and the adjoining Mexican state of Chiapas, the upper part of the nonmarine strata include gypsum and salt (Sapper, 1937, p. 27), suggesting that marine Middle or Upper Jurassic underlies northern Guatemala and Yucatán. There is a still undocumented report of marine Upper Jurassic in Honduras (Imlay, 1952, p. 969–970).

Marine Lower and Middle Jurassic, like marine Triassic, is unknown in the Caribbean region proper. Marine Lower Jurassic has been found at a locality along the south border of the region in Colombia: near the north end of the Cordillera Central of the Andes (Trumpy, 1943, p. 1299).

By middle Late Jurassic time (late Oxfordian; Imlay, 1952, p. 969) geosynclinial deposition reached the Greater Antilles, but only the westernmost island, Cuba. Thick marine deposits of that age, including rocks mildly metamorphosed to phyllite and quartzite, are found in westernmost Cuba. The much more strongly metamorphosed rocks of Isla de los Pinos (L. M. R. Rutten, 1934; Page and McAllister, 1944, p. 181–184), off the south coast of western Cuba, and the similar pre-ultramafic rocks of central and eastern Cuba, may possibly be of the same age. If they are, deformation and intrusion of diorite, now more or less gneissic, took place during a span from early Kimmeridgian to early Portlandian time, corresponding to a gap in the

Cuban sequence (Imlay, 1942, p. 1439–1440). Late Portlandian (Imlay, 1942, p. 1433–1440; 1952, p. 969) is represented in western, central, and eastern Cuba by 2000 feet of limestone, shale, chert, and some tuff.

Early Cretaceous

Early Cretaceous geosynclinal deposition took in all of Cuba and Hispaniola, probably Jamaica, and possibly Puerto Rico. If it included Jamaica, the deposits are represented in the metamorphic rocks of the Blue Mountains. This is the beginning of enormously thick volcanics: mostly pyroclastics in Cuba, mostly flows (including submarine flows) in Haiti, mostly pyroclastics in the Dominican Republic.

All the major subdivisions of the Lower Cretaceous are reported to be represented in Cuba (MacGillavry, 1937, p. 7–13; Maync, 1949, p. 531–532), but no records of a well-dated sequence have so far been published. Both the late Portlandian and the earliest Lower Cretaceous (Neocomian) of Cuba—the Aptychi limestones or Aptychi formation of Dutch geologists (MacGillavry, 1937, p. 7–8; M. G. Rutten, 1936, p. 10–12; van Wessem, 1943, p. 6–8)—contain concentrations of ammonite opercula, radiolaria, and pelagic foraminifera. They evidently are fairly deep-water deposits. The presence of land-plant debris is not opposed to that interpretation. Great quantities of decomposed land-plant debris were dredged by the Albatross in deep water off the Pacific coast of Central America (Agassiz, 1892, p. 11–12). The Cuban deposits are typically geosynclinal, typically eugeosynclinal, if that term is preferred.

The Lower Cretaceous of most of Cuba is made up principally of marine pyroclastics, with which minor flows are interbedded: part of the Tuff formation or Tuff series of Dutch geologists (de Vletter, 1946, p. 11–15; Hermes, 1945, p. 3–5; MacGillavry, 1937, p. 8–10; M. G. Rutten, 1936, p. 7–10; Thiadens, 1937, p. 11–18; van Wessem, 1943, p. 8–12; Vermunt, 1937, p. 15–17). Along the north coast of the island, in the northwestern part of Camagüay Province, van Wessem's (1943, p. 7–8) Aptychi formation includes thick beds of gypsum. Northward from the north coast of Cuba, in the subsurface strata of the Bahamas and peninsular Florida, the Lower Cretaceous consists of carbonate rocks. The only deep well so far drilled in the Bahamas penetrated almost 4000 feet of presumably Lower Cretaceous carbonate deposits and did not reach their base. Some of the dolomite of the carbonate deposits contains blebs of anyhydrite. The subsurface Lower Cretaceous carbonate rocks of peninsular Florida include minor amounts of bedded gypsum or anhydrite (P. L. and E. R. Applin, 1944, p. 1721).

The ultramafic rocks of Cuba probably were emplaced at an early date in Early Cretaceous time; that is, they probably are younger than the Central American ultramafics. In the east-central part of the island, where the ultramafics have been mapped (Flint, de Albear, and Guild, 1948, p. 43–45), they engulf schist and gneiss of unknown age and are overlain by volcanics that appear to range in age from Early to Late Cretaceous, but the lower, presumably Early Cretaceous, part is not yet dated. The late Upper Cretaceous deposits contain serpentine debris (L. M. R. Rutten, 1940, p. 542).

Much of what has been assigned to the Lower Cretaceous in Haiti, with varying degrees of qualification (Woodring, Brown, and Burbank, 1924, p. 86–93), is now known to be Upper Cretaceous. Nevertheless some of the lavas, particularly the earlier basalt of the northern part of the country and the basalt of the southern peninsula, may be Lower Cretaceous. Limestone between pillows of basalt, in the extensive area of basalt southwest of Port-au-Prince, shows sections of rudists comparable to those from Lower Cretaceous limestone in the Dominican Republic.

The metamorphic rocks of the Cordillera Central of the Dominican Republic have been claimed to be pre-Cretaceous, possibly Paleozoic (Weyl, 1940, p. 5, 28–29; and other publications by the same author). In the only area that has been mapped, however, a great thickness of sericite schist, altered tuff, and greenstone—all estimated to represent a thickness of 20,000 feet, if there is no duplication—overlies more or less altered limestone containing rudists and other fossils of Cretaceous age, probably late Early Cretaceous (Koschmann and Gordon, 1950 [1951], p. 317–325). Ultramafic rocks form a sill-like mass, between altered tuff and green-

stone in the upper part of the sequence. Limestone 70 miles southeast of the mapped area contains the rudists found in the mapped area and also *Orbitolina concava texana*, which occurs in the Aptian-Albian of Cuba, northern Central America, Colombia, and Venezuela.

Matley (1929, p. 465; 1951, p. 22) was convinced that the metamorphic rocks of the Blue Mountains in eastern Jamaica are pre-Cretaceous. They are now thought, however, to be Cretaceous or possibly Jurassic (Zans, 1953, p. 2). If they are Cretaceous, they are likely to be Lower Cretaceous.

Late Cretaceous

Late Cretaceous geosynclinal deposition spread still farther eastward to include all of the Greater Antilles, the Virgin Islands, and St. Croix. Figure 1 is a representation of Late Cretaceous paleogeography. Like most paleogeographic maps, it takes in too great a time span and therefore is composite.

Throughout the Greater Antilles, the Virgin Islands, and St. Croix the Upper Cretaceous deposits include basaltic and andesitic pyroclastics and flows of great thickness. They are everywhere strongly deformed and practically everywhere are intruded by batholiths of quartz diorite or granodiorite, near some of which they are altered to schist. It was on St. Thomas, one of the Virgin Islands, where the Swedish naturalist Cleve (1871, p. 5) was able to show almost 85 years ago that the metamorphic rocks are altered Cretaceous.

The latest dated Upper Cretaceous of the Greater Antilles includes, or consists mostly of, detrital deposits of late Campanian and Maestrichtian age (Imlay, 1944, p. 1013–1015). They include thin beds of limestone containing a distinctive fauna, characterized by the remarkable rudist *Barrettia*, which has been found in Cuba, Jamaica, Haiti, Puerto Rico, and northern Guatemala.

Deformation and intrusion of quartz diorite or granodiorite took place in the Greater Antilles and the Virgin Islands toward the close of Cretaceous time. The deformation was the first of several episodes of varying intensity during the interval from late Cretaceous to Eocene. In Cuba quartz diorite intrudes volcanics of early Late Cretaceous age, apparently Cenomanian to Coniacian (part of the Tuff formation or Tuff series of Dutch geologists; Imlay, 1944, p. 1011–1012) and contributed debris to the Campanian-Maestrichtian detrital deposits. The deformation and intrusion therefore is middle Late Cretaceous, approximately Santonian (Imlay, 1944, p. 1012). This deformation produced the arc of islands shown on Figure 1 where Cuba now is located. Discordance between the volcanics and the detrital deposits, however, is by no means uniform. At some localities there is no appreciable discordance, but the volcanics, the quartz diorite, and the ultramafic rocks were shedding debris into the late Campanian-Maestrichtian sea.

In Jamaica granodiorite intrudes the still undated metamorphic rocks of the Blue Mountains. Pebbles of granodiorite, however, are found in conglomerates of early Eocene age, but not in conglomerates interbedded with Upper Cretaceous limestone (Zans, 1953, p. 3). In Hispaniola, Puerto Rico, and the Virgin Islands deformation and plutonic intrusion took place at about the end of the Cretaceous, but the dating is not within narrow limits.

Altered basaltic and andesitic flows and tuffs, intruded by diorite, form the pre-Tertiary basement in Panamá (Woodring and Thompson, 1949, p. 227). The tuffs, like those in the Antilles, contain radiolaria and are altered to schist near some of the stocks of diorite. These volcanic rocks are shown as Tertiary volcanics on the current edition of the geological map of North America. At least part of the volcanics is dated as Late Cretaceous, on the basis of formanifera found in siliceous limestone in northwestern Panamá near the Costa Rican border.

Though the little Dutch islands off the coast of Venezuela (Aruba, Curaçao, Bonaire) are separated from the Greater Antilles by the width of the Caribbean Sea (a distance of 400 miles), their geology is Antillean. Their basement consists of a thick pile of volcanics overlain by detrital Upper Cretaceous (Molengraaff, 1929, p. 15–25; Pijpers, 1933, p. 7–38; Westermann, 1932, p. 11–37). Tuffs and cherts in the volcanic sequence contain radiolaria, as in the Greater Antilles. The three islands show

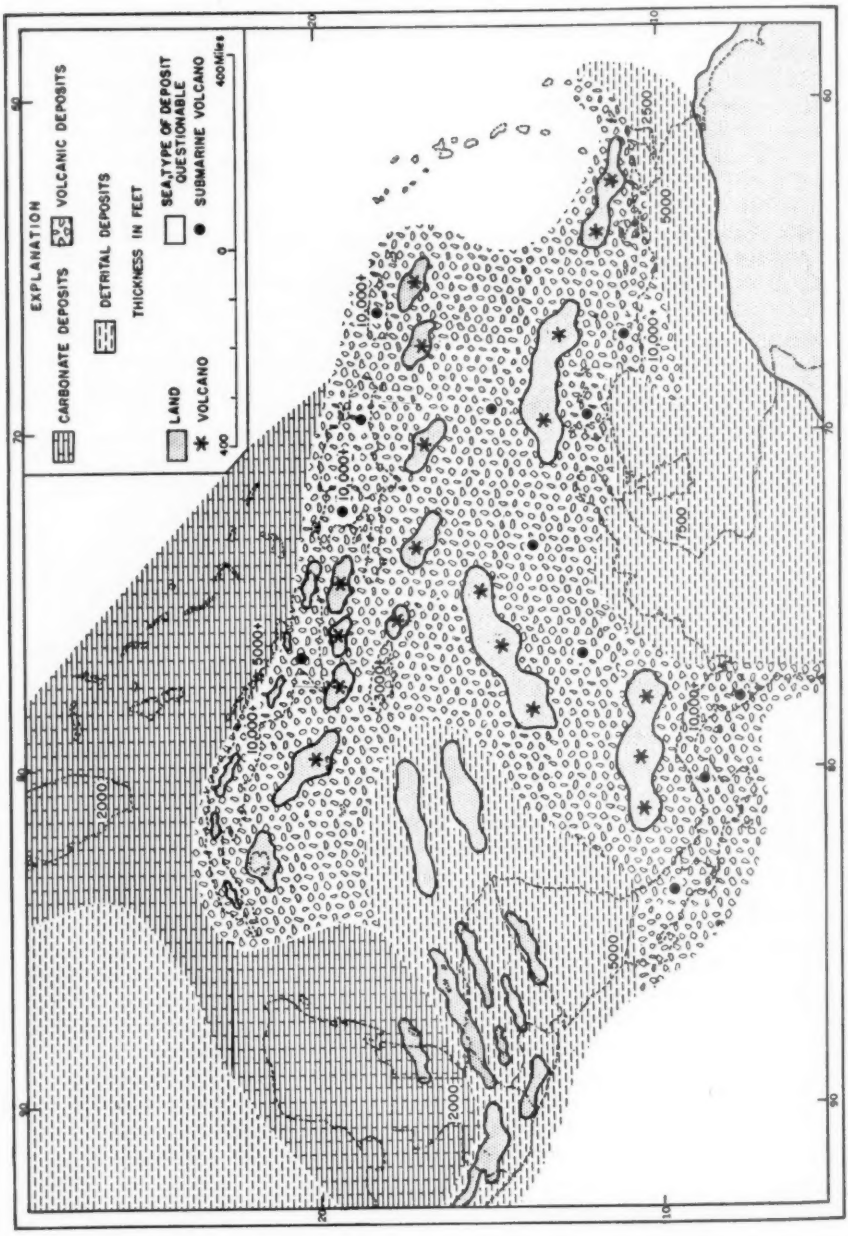

FIGURE 1.—CARIBBEAN REGION DURING LATE CRETACEOUS TIME
Present land indicated by dashed lines

a westward increase in intensity of deformation, in the size of exposed stocks of quartz diorite, and in degree of metamorphism.

The highly metamorphosed rocks of the Cordillera de la Costa in northern Venezuela (Bucher, 1952, p. 39–62, 73–84) include altered volcanics and apparently represent a transition from the volcanic sequence of the Dutch islands to the nonvolcanic sequence farther south and southwest in Venezuela. Late Cretaceous (Turonian) foraminifera have been found in limestone interbedded with schist, and the entire sequence of metamorphic rocks is now thought to be of Cretaceous age, or possibly both Cretaceous and Late Jurassic. Both Late Jurassic and Cretaceous fossils have been found in the metamorphic rocks of the Northern Range of Trinidad (Sutter, 1951–52, p. 189–191, 1951), the eastern continuation of the Cordillera de la Costa.

Where was the source of this great volume of Cretaceous volcanics along the north, west, and south borders of the Caribbean Sea? Though data are too meager for estimates, pyroclastics evidently are thicker and more widespread than flows. There doubtless were some minor volcanic centers in the present Greater Antilles, but they apparently were submarine. A northern source for the pyroclastics is at once eliminated. The entire Cretaceous sequence in the Bahamas and peninsular Florida, revealed by deep wells, consists of carbonate deposits. In north-central Cuba, the widespread Upper Cretaceous volcanic facies is replaced northward by a carbonate facies (de Albear, 1947, abstract; M. G. Rutten, 1936, p. 21–24). The only source for the pyroclastics is not in sight, it is under the waters of the present Caribbean Sea. Where the volcano-bearing lands were located, what their trend was, and whether they were large or small are matters that cannot be decided on purely geological grounds and may remain speculative for an indefinite time. The representation of the volcano-bearing lands on Figure 1 is one of several possible intrpretations.

FIGURE 1.—CARIBBEAN REGION DURING LATE CRETACEOUS TIME
Present land indicated by dashed lines

Paleocene and early Eocene

Paleocene deposits are mostly thin and of limited extent. In western Cuba the Paleocene is detrital (Bermúdez, 1950, p. 218–225), in southeastern Cuba volcanic (Lewis and Straczek, In preparation), in Haiti carbonate (Butterlin, 1953, p. 285). No Paleocene has been recognized so far in Jamaica or Puerto Rico.

Lower Eocene deposits, like the Paleocene, are mostly thin and not widely distributed. The lower Eocene of western and central Cuba consists of detrital and carbonate deposits (Bermúdez, 1950, p. 227–238), but in southeastern Cuba the lower Eocene is made up of volcanics, with which some limestone is interbedded (Bermúdez, 1950, p. 238–239; Lewis and Straczek). Carbonate deposits are the prevailing type in Hispaniola (Bermúdez, 1949, p. 19, 32; Butterlin, 1953, p. 285). The lower Eocene of Jamaica consists chiefly of detrital marginal marine-nonmarine deposits, and includes gypsum (Zans, 1953, p. 3). Land was near by, and the distribution of coarse detrital material suggests a source to the northeast (Trechmann, 1924, p. 4).

Strong deformation, stronger than the middle Late Cretaceous deformation, took place in western Cuba in early Eocene time (Palmer, 1934, p. 132, 141); none whatever in southeastern Cuba (Lewis and Straczek). In eastern Jamaica lower Eocene deposits are involved in intense folding (Trechmann, 1924, p. 5; 1929, p. 484; Zans, 1953, p. 3).

Middle and late Eocene

Carbonate deposits of middle and late Eocene age are widespread in the Greater Antilles and northward (Fig. 2). Both middle and late Eocene are strongly transgressive, particularly late Eocene. The carbonate deposits are thin in Cuba (Bermúdez, 1950, p. 239–263), moderately thick in Jamaica (Zans, 1953, p. 4), very thick in Hispaniola (Bermúdez, 1949, p. 10–11, 19–21, 32; Butterlin, 1953, p. 285; Woodring, Brown, and Burbank, 1924, p. 99–145[1]). If represented in Puerto Rico, they are overlapped.

The volcanic islands of the Late Cretaceous (Fig. 1) have disappeared, with the exception of remnants south of Cuba, where the deep

[1] The upper Eocene of this publication has been found, during recent years, to include middle and even lower Eocene.

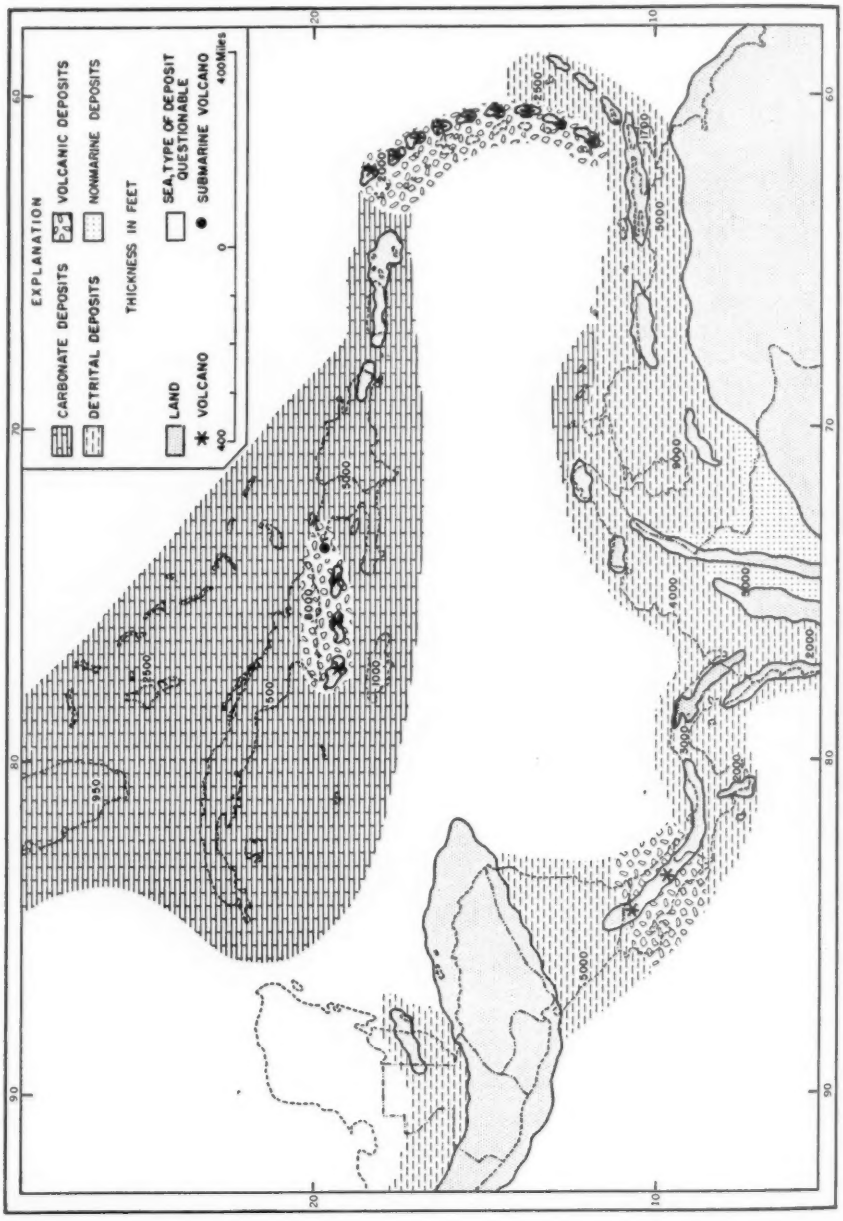

FIGURE 2.—CARIBBEAN REGION DURING MIDDLE AND LATE EOCENE TIME
Present land indicated by dashed lines

Bartlett Trough is now located. In southeastern Cuba the entire Paleocene and lower and middle Eocene are almost wholly volcanic—pyroclastics and relatively minor flows—representing an estimated thickness of 13,000 feet (Bermúdez, 1950, p. 238–239, 245–246; Lewis and Straczek; Taber, 1934, p. 576–581; Woodring and Daviess, 1944, p. 363–374[2]). The upper Eocene is mostly detrital (Bermúdez, 1950, p. 251–258; Lewis and Straczek; Taber, 1934, p. 584–585; Woodring and Daviess, 1944, p. 374–375). The Paleocene and Eocene pyroclastics are coarser southward, and flows are more numerous in the same direction, indicating a southward source. In northwestern Haiti pillow basalts are interbedded with limestone of middle or late Eocene age. It is fashionable—or perhaps was fashionable—to name hypothetical lands, despite the frequency with which some of these lands become unfashionable. The hypothetical land south of the Sierra Maestra of Cuba has been casually named Bartlett Land (Keijzer, 1945, p. 13, 86).

The middle Eocene of Jamaica is the Yellow Limestone of British geologists (Trechmann, 1923, p. 337–347). The fresh limestone is dark gray and contains small cubes of pyrite, oxidation of which produces the "yellow" color. Lithologically and faunally similar limestone, the Plaisance limestone, is found in northern Haiti (Woodring, Brown, and Burbank, 1924, p. 99–107). Elsewhere in Haiti middle Eocene limestone is ordinary white limestone, like the overlying upper Eocene limestone. The Cretaceous and Tertiary geology of Jamaica and northern Haiti, not southern Haiti, are similar in other features. This similarity does not support the recent proposition that Jamaica and southern Haiti have been moved 500 miles eastward to their present position along a hypothetical left-lateral fault (Hess and Maxwell, 1953, Fig. 2).

The volcanic foundation of at least the northern part of the outer sedimentary arc of the Lesser Antilles apparently was built during middle Eocene time, presumably from submarine volcanoes that later became volcanic islets, located immediately east of the arc. The French island of St. Bartholomew, near the north end of the sedimentary arc, has a foundation of andesitic pyroclastics, with which limestone of middle Eocene age is interbedded (Christman, 1953, p. 67–73). The volcanics are intruded by quartz diorite. The quartz diorite of the Lesser Antilles, heretofore recorded only as blocks in tuffs of middle-Tertiary age in Guadeloupe and Martinique, was formerly thought to indicate an Upper Cretaceous foundation for the sedimentary arc of the Lesser Antilles (Woodring, 1929, p. 416).

The detrital lower and middle Eocene of Barbados thickens southeastward and therefore suggests a source in that direction (Senn, 1940, p. 1557). Metamorphic rocks like the Cretaceous(?) metamorphics of Tobago would be suitable source rocks (Senn, 1940, p. 1561–1562)

Early Oligocene

Lower Oligocene is much more restricted in distribution than upper Eocene. Where represented in the Antilles, the lower Oligocene consists of carbonate deposits. They are thin in Cuba (Bermúdez, 1950, p. 264–270) and moderately thick in Hispaniola (Bermúdez, 1949, p. 21–23; Butterlin, 1953, p. 286). Though the White Limestone of Jamica ranges in age from late Eocene (or possibly middle Eocene) to early Miocene, an early Oligocene part has not yet been faunally defined.

Late Oligocene and early Miocene

Upper Oligocene and lower Miocene strata are widespread and strongly transgressive. Carbonate deposits are widely distributed (Fig. 3).

Deformation of varying intensity took place in some areas in the Greater Antilles toward the close of Oligocene time, and in southern Central America toward the close of early Miocene time. The thin early Miocene Güines limestone, the most widespread formation in Cuba (Bermúdez, 1950, p. 293–295; Palmer, 1934, p. 134–135; 1945, p. 17–18), was deposited after this period of deformation. Locally it overlies upper Oligocene marl and limestone without discordance, but it is widely transgressive. The anticlinal fold of the Sierra Maestra of southeastern Cuba, only the north limb of which is now visible, was

FIGURE 2.—CARIBBEAN REGION DURING MIDDLE AND LATE EOCENE TIME
Present land indicated by dashed lines

[2] The part of the Cobre formation covered by this publication and the Charco Redondo limestone member are now assigned to the middle Eocene, not upper Eocene.

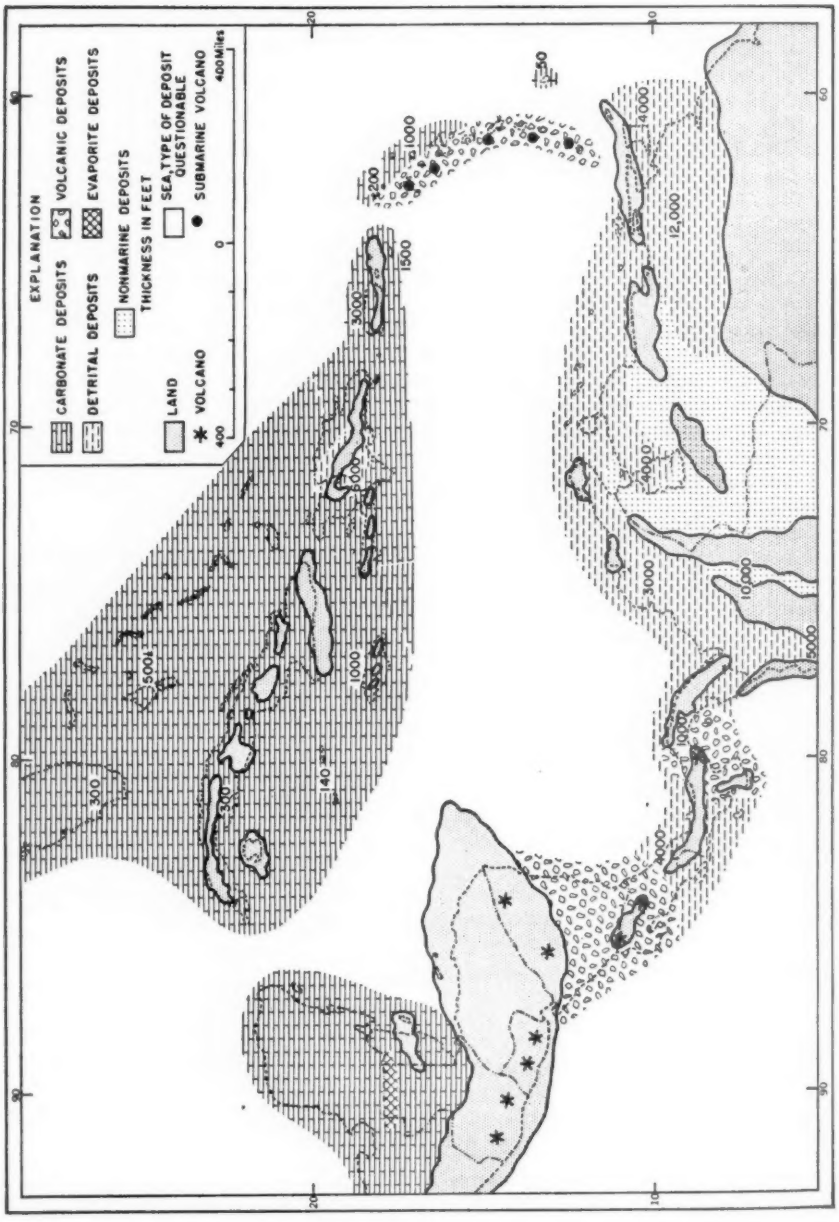

FIGURE 3.—CARIBBEAN REGION DURING LATE OLIGOCENE AND EARLY MIOCENE TIME
Present land indicated by dashed line

formed sometime during the Oligocene. By early or middle Miocene time (the dating depending on the age of the La Cruz formation), the anticline was disrupted by faulting. The thin coastal carbonate deposits of the La Cruz formation rest with marked unconformity on Eocene volcanics (Lewis and Straczek; Woodring and Daviess, 1944, p. 381–382). A fault along the southeast coast of Cuba, long advocated, therefore is supported by observable geological evidence. This fault has been interpreted as an extensive left-lateral fault (Hess, 1938, p. 86–87, fig. 6; Bucher, 1947, p. 109–111).

The Cordillera Central of the Dominican Republic and its continuation in Haiti, the Massif du Nord, were deformed, or at least uplifted, at approximately the end of the Oligocene. Along the borders of the newly emerged land the lower Miocene deposits are detrital, in contrast with the Eocene and Oligocene carbonate rocks (Bermúdez, 1949, p. 13–14, 27–29, 33–34; Woodring, Brown, and Burbank, 1924, p. 161–206). Marked discordance between upper Oligocence and lower Miocene, however, is rare. Farther from the newly emerged land the lower Miocene is made up of detrital and carbonate deposits.

Perhaps the foundation of the volcanic inner arc of the Lesser Antilles began to grow from submarine volcanoes in middle Tertiary time.

Middle and late Miocene

Hispaniola is the only Antillean island that has thick middle and upper Miocene strata; middle Miocene are more extensive than upper Miocene (Bermúdez, 1949, p. 14–17, 29–31, 35–36; Woodring, Brown, and Burbank, 1924, p. 219–223). Detrital deposits are the prevailing type, but carbonate and evaporite deposits are represented.

With few exceptions, formations younger than early Miocene in Cuba, Jamaica, and Puerto Rico are thin and marginal. Also, with few exceptions, they are mildly deformed or practically undeformed. In fact, the lower Miocene itself, again with few exceptions, is only mildly deformed, and in Puerto Rico no deposits of Tertiary age are strongly deformed.

Pliocene

The marine Pliocene deposits of the Caribbean region and its borders are of very limited extent and for the most part are thin. At the present time two divisions can be recognized: lower and upper. The lower Pliocene is represented, for example, by the Cabo Blanco formation of Venezuela, by marine strata near Cumaná, Venezuela, and by the Matura formation of Trinidad; the upper Pliocene by marine coralliferous strata at Limón, Costa Rica. In Haiti the marine deposits referred to the Pliocene (Butterlin, 1953, p. 286; Woodring, Brown, and Burbank, 1924, p. 239–243) may prove to be of late Miocene age, or to include upper Miocene, when the faunas are adequately studied.

The deformation at about the end of the Oligocene and at about the end of the early Miocene heralded stronger deformation that reached a climax during the Pliocene and during the Pleistocene, apparently early in the Pleistocene.

Pleistocene

Thin marine deposits on coastal terraces assigned to the Pleistocene are widely distributed. They may be almost entirely of late Pleistocene age. At the present time, however, there is no satisfactory basis for differentiating early Pleistocene, late Pleistocene, and Recent in the Caribbean region on a faunal basis, despite the availablity of large faunas at numerous localities. Reef corals are as abundant in the deposits referred to the Pleistocene as in the present meager reefs. In areas at the far margin of the Caribbean region, like the Bahamas, or at the far margin of the Antillean faunal province, like Bermuda, restriction of range affords a basis for differentiating Pleistocene from Recent. Restriction of range in Bermuda is illustrated by the Antillean terrestrial hermit crab (*Cenobita diogenes*). Throughout the Antilles the preferred shelter of this hermit crab is the large heavy shell of *Livona pica*. That species of marine snails lived in Bermudan waters during Pleistocene time but is now extinct there. In Bermuda the hermit crab seeks out Pleistocene

FIGURE 3.—CARIBBEAN REGION DURING LATE OLIGOCENE AND EARLY MIOCENE TIME
Present land indicated by dashed line

specimens of *Livona pica* to use as a shelter (Haas, 1950).

According to current age assignments, during Pleistocene time the present island of Hispaniola was divided into two islands by a strait along the Cul de Sac-Lago Enriquillo trough. The practically complete submergence of Cuba, shown by Schuchert (1935, Pl. 16), is without foundation.

Marine deposits assigned to the Pleistocene reach great altitudes in southeastern Cuba (700 feet), northwestern Haiti (1300 feet), and Barbados (1000 feet).

Unresolved Problems

Much remains to be done in the Caribbean region in an attempt to resolve problems, some of which involve fundamental matters in geology. First of all, detailed mapping is essential, particularly in pre-Tertiary areas, such as has been undertaken recently along the south border (Dengo, 1953; Smith, 1953). For the establishment of a standard Cretaceous sequence, Cuba offers great possibilities. A standard Cretaceous sequence and more data on the Jurassic sequence in Cuba are needed to determine the relations in space and time of the volcanic and plutonic rocks. There appear to be five periods of intrusion of granitic and dioritic rocks, ranging from late Paleozoic to early Miocene, including well-dated late Oligocene-early Miocene quartz diorite in Panamá and quartz diorite of probable Miocene age in Haiti. Absolute dating of these plutonic rocks would fix an upper limit for the age of the invaded sediments and would furnish data on the age of the metamorphic rocks. Is the apparent eastward spreading of geosynclinal deposition real, or is it an illusion resulting from inadequate dating?

What is the foundation of the Nicaraguan Swell, stretching from northern Nicaragua and southern Honduras to Jamaica? Why are the folds of southern Central America practically at right angles to those of northern Central America? What does the sigmoid curve of Panamá mean? It does not help much to call it an isthmian link and to attribute it to subsidence of the adjoining oceanic basins on both sides. To be sure, it is in fact an isthmian link. But why is it an outrageous isthmian link?

So much of the Caribbean region is under water that close co-operation between geologists and geophysicists is essential. What is the meaning of the anomlous combination of positive gravity anomalies and deep water in the Caribbean Sea and the Gulf of Mexico? The Puerto Rico Deep at the north end of the Vening Meinesz belt of strong negative anomalies has been found by Ewing and Worzel (1954, p. 168–169) to contain very thick unconsolidated sediments. At the south end the belt broadens to coincide with a basin containing 40,000 feet of Cretaceous and Tertiary sediments. In the intervening area, does the belt mark a sialic downbuckle, as postulated for the entire belt by Vening Meinesz and as elaborated by Hess? If the strong negative anomalies are due to great thickness of sediments along the length of the belt, what produced the long deep arcuate trough in which the sediments accumulated?

It is quite evident on ordinary geological grounds that a considerable part of the Caribbean Sea was land during the Cretaceous, and that at least some of this land was still land during the Eocene. How extensive was the land and how long-enduring? Geophysical data may aid in resolving that question.

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EXTENSION OF YORKTOWN FORMATION (MIOCENE) OF VIRGINIA INTO MARYLAND

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Abstract

In northeastern Tidewater Virginia and in southern Maryland the gray unweathered marine marls, clays, and sands of the Chesapeake group (Miocene) exposed in the high bluffs on Rappahannock and Potomac rivers and on Chesapeake Bay are overlain by 60–80 feet or more of sand and gravel. Previous investigators have treated this sand and gravel as one unit and referred it either to the so-called Lafayette formation (Pliocene?) or to the Sunderland formation (Pleistocene). We divide the sand and gravel unit into two formations.

The lower formation consists mainly of 40–60 feet of strikingly soft, loose, medium to coarse sand with interbedded lenses of gravel and subordinate light-gray clay partings. Glauconite is present in some beds. The basal bed is a fine gravel that rests unconformably on the unweathered Miocene strata, transgressing northwestward across both the St. Marys and the Choptank formations. Locally the materials are indurated to ferruginous sandstone and conglomerate, and one such mass at Nomini Cliffs (Horsehead Cliffs), Virginia, yielded an impression of *Halymenites major* Lesquereux. This fossil, representing a supposed marine organism, and the glauconite, the clay laminae, and the soft, loose texture of the sand, indicate the shallow marine origin of the formation.

Unconformably above the marine sand in the higher bluffs is 10–25 feet or more of medium to coarse, more or less ferruginous sand and lenses of gravel, with scattered pebbles and cobbles of quartz and occasional boulders of crystalline rock along the base (Pleistocene terrace deposits). These sands and gravels are locally indurated with iron oxide cement. In the higher bluffs (120–160 feet high) these deposits are interpreted to be the Sunderland formation; in the lower bluff (94 feet high) south of Jones Point on Rappahannock River, Virginia, they may belong to the Wicomico formation.

Fossils have been found in the Yorktown formation within a few miles of the lower sand and gravel formation, at altitudes that would permit the latter to fall within the stratigraphic limits of the Yorktown. The lower formation is therefore considered a northward extension of the Yorktown formation (Miocene).

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Introduction

This paper is a slightly modified and expanded version of a paper presented before Section E of the American Association for the Advancement of Science at Richmond, Virginia, in December 1938. An abstract was published at that time (Stephenson and MacNeil, 1938, p. 1959). The interpretation offered in the abstract apparently did not lead geologists to consider seriously the possible extension of the Yorktown formation from Virginia north-

ward into southern Maryland. It therefore seems desirable that the more detailed evidence presented in the completed paper be made available.

The Miocene sediments of Maryland heretofore recognized, provincially known as the Chesapeake group, consist of gray unweathered marine marls, clays and sands, some beds of which are abundantly fossiliferous. These beds have been differentiated in ascending order into the Calvert, Choptank, and St. Marys formations (Clark *et al.*, 1904). These formations extend southward into Virginia where, however, a still higher unit of the Miocene series, the Yorktown formation, is present.

The thicknesses of formations given are only approximate; however, errors of 2 or 3 feet are not sufficient to affect the conclusions.

Description, Relations, and Age of Sand and Gravel Unit

The Nomini Cliffs on Potomac River in Westmoreland County, Virginia (Fig. 1), form an east-west series of high north-facing bluffs about 5 miles long. In local usage the Cliffs are divided from east to west into the Nomini Cliffs proper, the Stratford Cliffs, and the Horsehead Cliffs. The geologic units exposed in this series of cliffs are indicated in published columnar sections (Stephenson, Cooke, and Mansfield, 1933, p. 27, Fig. 7). These units as then understood included in ascending order the Calvert, Choptank, and St. Marys formations of the Miocene series, totaling 60–90 feet of strata, overlain by 60–80 feet of sand with subordinate beds of gravel, designated (in the illustration cited) as late Miocene or Pliocene and Pleistocene, undifferentiated. From east to west the sand and gravel unit transgresses the St. Marys formation and in the Horsehead Cliffs rests directly on the Choptank formation. It is to this sand and gravel unit that attention is called in this paper.

In the Nomini Folio, Darton (1896), classified the undivided unit as the Lafayette formation of W J McGee (1891), and ascribed to it a marine origin. Clark and Miller (1912, p. 129–131) correlated the same unit with the Sunderland formation (Pleistocene) of Maryland. Mansfield (1926, p. 177) referred this part of the section to the Pleistocene but did not assign it to any formation.

We believe that the undifferentiated body of sand and gravel constituting the upper part of

Figure 1.—Sketch Map Showing Approximate Locations of Sections 1–8 of Figure 2

the section in the Nomini Cliffs (Fig. 2, sec. 6) is separable into two formations. The lower formation consists mainly of 40–60 feet of strikingly soft, loose, medium to coarse, sparingly glauconitic sand with interbedded lenses of fine gravel, and subordinate light-gray clay partings; the basal bed consists of 2–3 feet of fine gravel resting on gray unweathered sands and clays referable, in places, to the St. Marys formation and elsewhere to the Choptank formation, of the Miocene. The sands are cross-bedded and are locally and irregularly indurated to ferruginous sandstone. In one such mass of sandstone at Horsehead Cliffs are imprints of the problematical organism *Halymenites major* Lesquereux (Pl. 1, fig. 2). The presence of this organism, the soft texture of the sand, the sweeping lines of cross-bedding, and the glauconite grains, indicate deposition in shallow marine water. The gravel at the base of the sand, the fact that the base of the gravel in the Stratford Cliffs cuts across a thickness of 5 or 6 feet of the underlying nearly horizontal strata, and

FIGURE 2.—SECTIONS SHOWING SUGGESTED CORRELATION OF LOWER SAND AND GRAVEL FORMATIONS AT NOMINI CLIFFS, VIRGINIA, AND IN CALVERT COUNTY, MARYLAND, WITH YORKTOWN FORMATION OF VIRGINIA

Differences in hypsographic position of given formations with respect to sea level in different columns are due to differences in geographic position; sections showing low positions are down dip from those showing high positions.

the absence of the St. Marys formation in the Horsehead Cliffs indicate that an erosional unconformity separates the loose sand unit from the St. Marys and Choptank formations.

The upper of the two sand and gravel formations consists of 10–25 feet or more of coarse reddish pebbly sand, with interbedded gravel lenses; the base is marked by a gravel bed of varying thickness, in which are sparse smoothly water-worn cobbles up to 10 inches long. This unit is interpreted as a Pleistocene terrace deposit, and on the basis of its surface altitude of about 160 feet above sea level, a short distance south of the edge of the cliff, we correlate it with the Sunderland formation of Maryland.

Mansfield (*in* Stephenson *et al.*, 1933, p. 28) has shown that the beds of the St. Marys formation beneath the lower sand and gravel formation in the Nomini Cliffs represent only the lower part of that unit; the upper part, and in places all, of the St. Marys has been removed by erosion. The loose sand and gravel unit must therefore be younger than the lower beds of the St. Marys formation and older than the Sunderland formation.

At a bluff about 94 feet high on Rappahannock River, half a mile south of Jones Point (Fig. 2, sec. 5), north of the Middlesex County line (near Butylo), 30 miles south-southeast of Nomini Cliffs, a bed of shell marl within 10 feet of the base of the bluff (largely concealed at the time examined) contains *Pecten madisonius* Say, *Pedalion maxillatum* (Deshayes), *Mactra* sp., and other fossils that indicate its St. Marys age (Mansfield 1931, p. 6, and oral communication). An oxidized ferruginous zone about 60 feet above the beach yielded imprints of *Pecten jeffersonius* Say; Mansfield believed it represents the lower part (zone 1) of the Yorktown formation, the youngest of the Miocene formations of Virginia. This fossil-bearing bed is overlain by 24 feet of loose, fine to coarse sand with interbedded lenses and stringers of fine gravel; near the middle the sand is strongly glauconitic (confirmed by C. S. Ross, personal communication); in a ferruginous layer about 2 feet above the base a few indeterminate prints of fossil mollusks were observed. The contact of this loose sand and gravel unit with the underlying fossil bed was not clearly exposed. Beneath the fossil-bearing bed (zone 1, Yorktown) is 21 feet of loose, fine sand resting on light-gray clay of the St. Marys formation.

At the top of the bluff is 10 feet of coarse ferruginous sand and gravel with a few pebbles and cobbles up to 6 inches in length along the base; one basal boulder of coarsely crystalline pegmatite 15 inches in diameter was observed. This is a Pleistocene terrace deposit.

The 24-foot loose sand and gravel unit above the shell marl possesses the physical characteristics of the loose sand unit between the St. Marys formation and the terrace deposit correlated with the Sunderland formation in the Nomini Cliffs. At the cliff south of Jones Point (Fig. 2, sec. 5) the enclosing formations indicate that the loose sand and gravel unit is younger than basal Yorktown and older than the Pleistocene terrace deposit.

On Rappahannock River about 1¼ miles south by east of the preceding bluff, a quarter of a mile northwest of Bayport, the loose sand unit is well exposed in a bluff.

On U. S. Highway 360, on the north-facing slope of Piscataway Creek valley, 4¼ miles (airline) south-southwest of Tappahannock, Essex County (Fig. 2, sec. 4), a coarse ferruginous sandstone about 2 feet thick crops out in the slope about 90 feet above sea level and is underlain by several feet of loose greenish-gray glauconitic sand, like that above zone 1 of the Yorktown in the bluff south of Jones Point. Of particular interest is the presence of many imprints of *Halymenites major* in the ferruginous sandstone (Pl. 1, fig. 1) like the imprints of this organism in a sandstone mass

PLATE 1.—CASTS FROM IMPRINTS OF *HALYMENITES MAJOR* LESQUEREUX

FIGURE 1.—VIEW OF RUBBER CAST OF IMPRINT (× 1)

From Yorktown formation, 4¼ miles south-southwest of Tappahannock, Essex County, Virginia.

FIGURE 2.—VIEW OF PLASTER CAST OF IMPRINT (× 1)

In upper part of sand and gravel unit (Yorktown formation) at Horsehead Cliff, Westmoreland County, Virginia.

FIGURE 1

FIGURE 2

CASTS FROM IMPRINTS OF *HALYMENITES MAJOR* LESQUEREUX

of identical lithologic character in the lower of the two loose sand units at Nomini Cliffs in Westmoreland County (Pl. 1, fig. 2).

Fossils collected in King and Queen County (Fig. 2, sec. 3), about a mile southwest of Millers Tavern, 9 miles southwest of Tappahannock, were identified by Doctor Mansfield as belonging to a *Chama*-bearing bed (zone 2, Yorktown formation) lying above zone 1, on James River near Grove Wharf in James City County; the bed near Millers Tavern is about 140 feet above sea level. Mansfield also identified fossils representing the same bed from a locality in the same county half a mile southwest of Powcan (Fig. 2, sec. 2), 12 miles southwest of Tappahannock, where the bed is estimated to lie 120 or 130 feet above sea level. At Watkins Mill in King and Queen County (Fig. 2, sec. 1), 4 miles southwest of Millers Tavern, Mansfield observed a conglomeratic bed at the base of zone 1 of the Yorktown formation, the altitude of which he estimated to be about 95 feet above sea level; this is believed to be a basal conglomerate of the Yorktown formation, and it probably corresponds to the fine gravel bed at the base of the loose sand and gravel formation at Nomini Cliffs.

The stratigraphic relation of the fossil-bearing beds in the lower part of the Yorktown formation near Millers Tavern and near Powcan to the loose sand and gravel formation was not determined with certainty, but, in view of the flat-lying attitude of the Miocene strata in this part of Virginia, the sand and gravel unit is believed to be a shallow-water, near-shore facies of the Yorktown formation.

In a cut on U. S. Highway 360 2 miles west by north of Warsaw, Richmond County, 8 or 10 feet of loose sand with light-gray clay partings belonging to the loose sand and gravel unit is exposed beneath 27 feet of coarse pebbly sand, a Pleistocene terrace deposit.

The loose sand and gravel unit is not restricted to Virginia but extends northeast into Southern Maryland. It is well developed in the upper part of a series of high cliffs facing Chesapeake Bay between Little Cove Point and Drum Point in southern Calvert County (Fig. 2, sec. 7). Here the unit overlies 55 or 60 feet of the St. Marys formation, some parts of which are richly fossiliferous. The sand and gravel unit consists of 30–55 feet of fine to coarse and pebbly, loose sand with interbedded layers of fine gravel and with a bed of fine gravel of varying thickness along the base; interbedded light-gray clay films and thin layers are locally present in subordinate amounts; locally the sand and gravel are indurated to ferruginous sandstone masses. In the higher bluffs the sand unit is overlain by 10–15 feet of Pleistocene terrace sand and gravel, the base of which is marked in places by a line of large smoothly water-worn cobble stones.

The sand and gravel unit is well exposed in the upper part of a high bluff 6 miles northwest of Cove Point, a mile south by east of Flagpond (Fig. 2, sec. 8). It consists of 15–25 feet of loose, fine to coarse, more or less pebbly sand with subordinate amounts of light-gray clay films and partings; a gravel bed at the base ranges from 3 to 10 feet in thickness. The unit rests with apparent unconformity on the St. Marys formation at an altitude of about 85 feet above sea level. Here as elsewhere the unit is overlain by 10–15 feet of coarse Pleistocene sand and gravel.

Summary

A body of sand and gravel in northeastern Tidewater Virginia and in southern Maryland, formerly correlated with the so-called Lafayette formation (Pliocene?) and by others with the Sunderland formation (Pleistocene), is divisible into two formations. The lower formation consists of 40–60 feet of loose sand and gravel resting unconformably upon, and transgressing across fossiliferous strata of the St. Marys and Choptank formations (both Miocene) in the Nomini Cliffs, Virginia, and farther south resting upon strata as young as early Yorktown; the presence of glauconite, of *Halymenites major* Lerquereux, and the soft texture of the sand, indicate a marine origin for this unit. The geographic position of this unit within a few miles of fossiliferous, lithologically closely similar beds of known Yorktown age at about the same altitude above sea level, in an area where the Miocene strata have a regional dip of probably less than 5 feet to the mile, justifies the conclusion that the unit is of Yorktown age. If this conclusion is correct, the Yorktown formation extends farther north than it has

previously been mapped and reaches Chesapeake Bay in the Calvert Cliffs in southern Calvert County, Maryland. This extension seems reasonable because the Yorktown is a transgressing formation and extends to the crystalline rocks of the Piedmont Plateau in the vicinity of Petersburg, Virginia.

The upper formation is terrace sand and gravel of Pleistocene age, ranging from 10–25 feet thick. It might seem that the underlying loose sand and gravel unit might represent an off-shore facies of the so-called Lafayette (Pliocene?) high-level alluvial gravels in the vicinity of Falls Church, Virginia, and in the high parts of the District of Columbia. However, reference of these high-level gravels to the Pliocene is not established, and they too may be late Miocene (Yorktown).

No fossiliferous beds of Pliocene age have been recorded from northeastern Tidewater Virginia or eastern Maryland. To provide an embayment into which marine Pliocene sediments could have been transported, at the same level as the known marine Miocene sediments in the area under consideration, it would be necessary to postulate either a shallow erosion basin or a basin produced by broad downwarping. There is no known evidence to support either postulate. The interpretation here presented—that the loose sand and gravel unit exposed in the Nomini Cliffs in Virginia and in the high cliffs along the southern 10 miles or more of the Calvert Cliffs in Maryland represents a northward extension of the closely similar loose sand and gravel facies of the Yorktown formation of Virginia—seems in accord with the known facts.

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CORRELATION OF COPPER MINERALIZATION WITH HYDROTHERMAL ALTERATION IN THE SANTA RITA PORPHYRY COPPER DEPOSIT, NEW MEXICO

By PAUL G. LEROY

ABSTRACT

The Santa Rita intrusive of Southern New Mexico consists of highly altered and strongly fractured granodiorite porphyry. Minor fracturing of the ingneous body is probably the combined effect of localized tectonic activity and stress developed in the crystallized shell of the cooling mass by continued movement of the magma. Previous studies indicate four stages in the hydrothermal alteration of the porphyry, starting with nearly fresh rock and ending with a highly siliceous and sericitized counterpart. Chemical changes of wall-rock alteration indicate a loss of Na_2O, CaO, MgO, and SiO_2, with a relative gain in water of hydration and sulfur in argillized rock. Loss of Na_2O, CaO, MgO, and FeO continues during sericitization, while potash, sulfur, and probably silica increase. Copper mineralization consists of disseminated supergene chalcocite and pyrite coated with chalcocite. Chalcocite also coats quartz-sericite-pyrite veinlets. Chalcopyrite and bornite are hypogene. Intergrowths of chalcopyrite in bornite suggest that some of the hypogene copper mineralization took place at 475° C or higher. Primary copper minerals seem to antedate the quartz sericitization of the Santa Rita porphyry. The concentration of supergene chalcocite in stage 3 seems to be due partly to the relatively greater porosity and higher pyrite content of this rock. The high porosity and pyrite content of sericitized porphyry may similarly render stage 4 favorable to economic chalcocite mineralization. Stage 3 or argillization of granodiorite appears to be the stage of alteration most favorable to ore, for, besides having a porosity and pyrite content high enough to facilitate noteworthy secondary enrichment, this stage of hydrothermal alteration also furnished the hypogene copper minerals which constitute the protore of the deposit.

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INTRODUCTION

Nature of Problem

Renewed investigation of the porphyry copper deposits of the western United States has recently developed because of increased attention given to wall-rock alteration and its relationship to metallic mineralization. This paper presents the results of a field and laboratory investigation devoted to the relationship between hydrothermal alteration and copper mineralization in the Santa Rita porphyry copper deposit of southern New Mexico.

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Professor Paul F. Kerr of Columbia University suggested the investigation as a sequel to his study, and he has given much advice both in the field and during laboratory research. The writer is indebted to him also for critical reading of the manuscript. Thanks are also due Professors Charles H. Behre, Jr., and Arie Poldervaart, both of Columbia University, who reviewed the paper and made helpful suggestions.

Location

The Santa Rita porphyry copper deposit in the Central Mining District of southwestern New Mexico approximately 150 miles northwest of El Paso, Texas (Fig. 1), is the easternmost mineralized porphyry exploited in a belt of scattered intrusives extending from northern Mexico to Montana. Within a radius of 300 miles lie San Manuel, Cananea, Morenci, Ray, Castle Dome, Miami, Globe, and Ajo, deposits of similar nature and of major importance. The Chino property is listed among the first six major domestic copper producers. A substantial zinc production has been developed from neighboring mines (Lasky and Hoagland, 1950).

FIGURE 1.—LOCATION MAP OF SANTA RITA PORPHYRY COPPER DEPOSIT, NEW MEXICO

Previous Geological Studies

The literature on the Santa Rita and surrounding areas has been previously summarized (Spencer and Paige, 1935; Kerr *et al.*, 1950). Papers outlining the general geology, salient features of the district, and which provide a background for the problem are by: Paige (1916); Spencer and Paige (1935); Schmitt (1935); and Kerr, Kulp, Patterson, and Wright (1950). The following recent papers on wall-rock alteration are also of value: Lovering (1941; 1949; 1950); Peterson *et al.* (1946); Schwartz (1947); Sales and Meyer (1948; 1950); Kerr *et al.* (1950); Kerr (1951). Early descriptions of other mining districts, especially Lindgren (1905); Ransome (1919); Reber (1916); and Spencer (1917) also furnished much information.

General Geology

General statement.—As detailed discussions are available, the general geology of the Santa Rita area is only briefly outlined. On the geologic map of New Mexico (Darton, 1929, 1/500,000), the Central Mining District corresponds roughly to a triangular area of Paleozoic and Mesozoic rocks surrounded by Tertiary lavas, tuffs, and sediments. The base of the triangle is approximately coincident with parallel 32°45′ N. Lat. and lies between meridians 107°50′ and 108°10′. Just west is the Silver City-Pinos Altos area which also exposes Paleozoic and Mesozoic formations.

Sedimentary rocks.—Near the Santa Rita intrusive Mississippian to Cretaceous rocks are found. Early Paleozoic formations outcrop to the north. On the margins of the intrusive the country rock has been intensely metamorphosed and identification of individual formations is uncertain. The sedimentary rock units reported by Kerr and his coworkers (1950, p. 284) during geological mapping of the pit area are listed in Table 1.

TABLE 1.—STRATIGRAPHY OF THE SANTA RITA PIT AREA

System	Formation	Approximate thickness in feet	Character of rocks
Cretaceous	Colorado	300–1500	Shales and sandstones and intercalated diorite sills
	Beartooth	40–135	Red quartzites, red and gray shales
Pennsylvanian	Syrena (Upper Magdalena)	360–390	Shaly limestone and shales

Igneous rocks.—Santa Rita granodiorite porphyry is the main igneous rock exposed in the open pit. The least-altered rocks are generally porphyritic, but equigranular facies of the same composition were also found (Fig. 8). Phenocrysts of plagioclase, biotite, quartz, and hornblende are set in a fine-grained groundmass of quartz and orthoclase, with magnetite, sphene, apatite, and zircon as accessory minerals.

Two of three quartz diorite sills, intercalated in the Colorado formation, are exposed in the Santa Rita mine. The lower sill, known in the Groundhog mine southwest of Santa Rita, thins out before reaching the south pit. The middle sill is 175 feet thick and is fine-grained and equigranular. The upper sill, 150 to 200 feet thick at Santa Rita, is porphyritic with plagioclase and doubly terminated quartz phenocrysts. Ferromagnesian minerals are typically altered to epidote and chlorite.

At least four types of dike rocks may be observed. Early aplite dikes cut the granodiorite porphyry in the north pit. Thin, nearly horizontal sheets of aplite are also common and may represent emplacement of a late crystallization product of the granodiorite magma along flat-lying joints developed in the roof of the cooling intrusive (Balk, 1937, p. 39). Two types of granodiorite dikes may be distinguished: one in which the dominant mafic mineral is hornblende and which contains few quartz phenocrysts, and another, later set of dikes in which quartz phenocrysts are abundant and which also carries large, pink, poikilitic orthoclase phenocrysts. The younger dikes are outlined along benches of the south pit by a green to rusty brown color. A highly bleached dike in the east wall of the south pit appears to be the altered equivalent of a hornblendic variety of granodiorite porphyry dike, suggesting that both types of dikes are later than the Santa Rita intrusive. Latite dikes are the youngest exposed intrusives and generally strike easterly, crosscutting the north-northeast striking granodiorite dikes. The latites are dark gray with lustrous biotite phenocrysts and little quartz.

Structure.—At least three periods of faulting seem represented: (1) after the emplacement of the quartz diorite sills and prior to the granodiorite dikes, (2) after the granodiorite dikes, and (3) post Tertiary volcanics. Most faults are normal, though Schmitt (1939, p. 786) reports a low-angle overthrust related to the Hanover intrusion. In places there is evidence of repeated movements along the same zones of weakness, and many of the faults displacing lavas and tuffs south of the Santa Rita pit reflect corresponding displacements in Mesozoic and earlier rocks beneath the volcanics. Away from the Santa Rita intrusive the beds dip gently southeast, but marginally dips steepen to 10°–55°, probably because of the intrusion of the Santa Rita porphyry.

Wall-rock alteration.—Pyrometasomatism has produced tactite and magnetite bodies adjacent to both the Hanover and Santa Rita intrusives. Contact metasomatic silicates and related zinc ore deposits are also developed on the margins of certain granodiorite dikes. Hydrothermal

alteration developed chlorite, clays, quartz, and sericite within the Santa Rita intrusive and in the adjacent sills. Kerr and his colleagues (1950, p. 310) recognize four stages of alteration of the granodiorite porphyry. Disseminated copper mineralization accompanied this widespread rock alteration.

Mineral deposits.—The Santa Rita porphyry yields principally chalcocite disseminated in the intrusive and in the mineralized portions of surrounding sills. A notable amount of molybdenite is also recovered in the treatment of low-grade copper ore. Though magnetite is abundant to the north of the Santa Rita pits, extraction has not been undertaken, in contrast to the magnetite bodies formerly mined near Fierro. The district is an active zinc producer. Sphalerite and subordinate galena are produced from both fissure fillings and limestone replacement deposits.

Structural Features of the Santa Rita Intrusive

General Statement

A complete treatment of the structural features of the Santa Rita porphyry intrusive was not the object of this investigation, though features closely related to the hydrothermal alteration and metallic mineralization of the porphyry were studied in detail.

Causes of Fracturing

Intense fracturing of the host rock is characteristic of the porphyry copper deposits. At Santa Rita the fracturing takes the form of innumerable joints, slickensided fault planes, cemented fractures, and a ramifying network of veinlets. The intensity and pervasiveness of shattering is demonstrated by repeated failures to chip off hand specimens of satisfactory size from an outcrop.

Fracturing of the altered igneous rocks has been attributed to various causes. Volume shrinkage due to crystallization and thermal contraction is postulated by Carlton D. Hulin (1948, p. 47). At Santa Rita some fracturing undoubtedly developed as a result of cooling and shrinkage in the peripheral zone of the intrusive, and the presence of flat-lying aplite bands is also suggestive. There is evidence that some of the barren quartz veinlets formed early in the history of the Santa Rita deposit and may thus be related to fracturing produced by shrinkage in the quickly cooled shell of the igneous mass.

Pennebaker (1942, p. 130) noted that not all the porphyry bodies at Ely were fractured, and he discounted shrinkage on cooling, suggesting tectonic activity instead. This is supported by the Chuquicamata deposit where the fractures were formed by strong, north-south shear zones with strong tension fractures within the zones (Lopez, 1942, p. 127). Anderson (1948, p. 177) draws attention to the location of the mineralized area at Bagdad, Arizona, at the intersection of two shear zones.

Locke (1933, p. 622) considers shrinkage during mineralization as the cause of widespread shattering of porphyry copper deposits. More recently Gilluly (1942, p. 272) suggested collapse by escape of volatiles for the fracturing of the Ajo copper deposit. Thus mineralization or the action of mineralizing fluids is added to the list of possible causes of the thorough fracturing of the porphyry copper deposits.

Regional structural control in the emplacement of the granodiorite and related igneous bodies of the Central Mining District is indicated by the concentration of intrusives, dikes, and faults in a belt roughly 2 miles wide trending northeast to north from the Ground Hog mine to Fierro (Spencer and Paige, 1935, Pl. 1). The regional pattern is evident even in the Tertiary, postvolcanic structural movements. The localization of the Santa Rita deposits at the intersection of two main structural zones has not been proved and possibly does not exist, but the persistence of tectonic activity in the Santa Rita area cannot be disregarded. However, there are indications that the intricate fracturing of the Santa Rita granodiorite did not all originate at the same time, but that the process extended over a considerable period. Some of the fracturing probably occurred soon after the crystallization of the intrusive, prior to its hydrothermal alteration, and may be the result of shrinkage upon rapid cooling. Apparently later fracturing permitted the rise of mineralizing fluids along definite channels in the porphyry following the

widespread hydrothermal alteration. The mechanics of intrusion may also have contributed to fracturing of the porphyry. In upper benches of the north wall of the south pit to large sedimentary xenoliths are exposed. Some of the sediments have been largely replaced by magnetite and quartz. These magnetite-bearing beds are cut by numerous granodiorite dikes, generally altered to stage 2. It is inconceivable that the dikes themselves caused the pyrometasomatism of the sediments for they contain inclusions of magnetite-bearing sediments (Fig. 2). Therefore, the sediments must have been pyrometasomatized earlier, probably when in contact with the rising intrusive. Renewed upward movement of the magma engulfed them in the porphyry and permitted small granodiorite dikes to penetrate along fractures. Selective metamorphism of the inclusions after intrusion is also possible but is regarded as unlikely. Thus repeated upward movement of the magma may have caused further fracturing of the crystallized marginal portions of the intrusive, providing passages for later hydrothermal alteration and mineralization of the fractured rock.

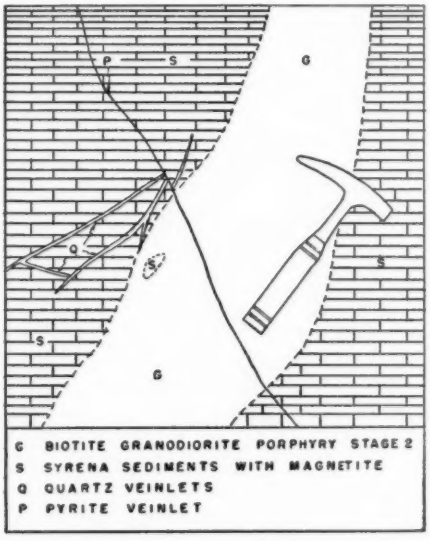

FIGURE 2.—GRANODIORITE PORPHYRY CUTTING ALTERED SYRENA SEDIMENTS

Rather than the result of a single process active during a limited time, minor fracturing of the porphyry copper deposits is probably the combined effect of several causes over a prolonged period.

Veinlets

Types of veinlets.—Within the more highly altered parts of the south pit, the granodiorite is traversed by an intricate network of veins ranging in thickness from a fraction of an inch to 2 inches or more. Three types of veinlets may be observed in the field: barren quartz veinlets, a quartz-alunite type, and a quartz-sericite-pyrite type.

The barren quartz veinlets are most numerous but also show the least continuity—*i.e.*, the quartz-filled fractures are offset within short distances by other quartz veinlets or, more commonly, by veins of the other two types.

The quartz-alunite veinlet consists of a light-greenish, fine-grained material. Alunite generally forms a central filling between two quartz zones. In some veinlets, usually an inch or more in thickness, alunite fills the space between the comb quartz which projects into the center of the vein from each wall. In contrast to the barren quartz veinlets, many of the quartz-alunite veinlets contain small molybdenite flakes at their margins.

The third type consists chiefly of quartz, sericite, and pyrite in order of abundance. The barren quartz and quartz-alunite veinlets have conspicuously sharp and straight contacts and are apparently simple cases of vein material filling narrow, open fractures, with almost negligible reaction with wall rock. The quartz-sericite-pyrite veinlets have irregular, gradational wall margins. For these, replacement of wall rock outward from a thin fissure appears to be the dominant process. The altered granodiorite is "flooded" by quartz-sericite on both sides of the pyrite seam. This alteration "halo" may vary from a fraction of an inch to an inch or more on either side. The pyrite-bearing veinlets tend to be of considerable economic importance, since much of the pyrite is coated or partially replaced by black chalcocite. In places a thin coat of greenish malachite stains

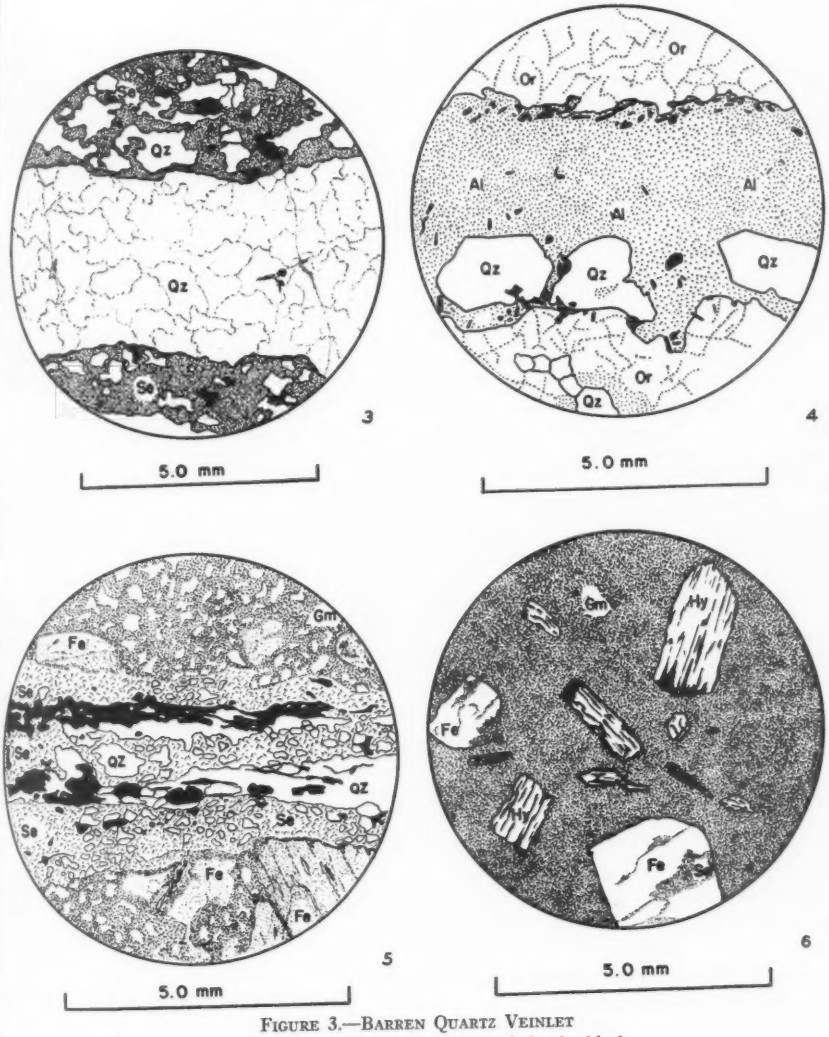

FIGURE 3.—BARREN QUARTZ VEINLET

Quartz (Qz), sericite (Se), pyrite and chalcocite black.

FIGURE 4.—QUARTZ-ALUNITE VEINLET

Quartz (Qz), alunite (Al), orthoclase partly altered to alunite (Or), molybdenite black.

FIGURE 5.—QUARTZ-SERICITE-PYRITE VEINLET

Quartz (Qz), feldspars altered to kaolinite and sericite (Fe), sericite (Se), groundmass of quartz and orthoclase (Gm), pyrite and chalcocite black.

FIGURE 6.—HORNBLENDE GRANODIORITE PORPHYRY DIKE

Hydromica after hornblende and biotite (Hy), feldspars altered to koalinite, sericite (Se), fine-grained groundmass of quartz, orthoclase and clay.

the chalcocite and renders the veinlet even more conspicuous in the bleached porphyry.

Microscopic description early barren quartz veinlets.—In the barren quartz veinlet (Fig. 3) the quartz occurs as irregularly shaped, interlocked grains, 0.1–1.2 mm diameter, averaging

about 0.6 mm diameter. Much of the quartz shows strain shadows under crossed nicols, and small fractures cut across two or more grains, indicating later deformation. Larger fractures transect the whole veinlet and allowed later infiltration and replacement of quartz by sericite. The quartz is dusty and under high power is literally riddled with inclusions—cavities nearly 0.01 mm across, empty or partly filled with liquid. Most inclusions are scattered through the quartz, but some follow a trend and cut across grain boundaries, suggesting a secondary origin. The latter may be related to the hydrothermal alteration of the granodiorite porphyry (Tuttle, 1949, p. 353). Some cavities are larger than average and besides bubble and liquid contain a sharply outlined transparent cube, 3 to 4 microns to the side. This cube is so small that it appears to have the anisotropism of the surrounding quartz. Despite its apparent double refraction the transparent cube is thought to be a common isometric salt, probably sodium chloride (Tuttle, 1949, p. 333). Similar inclusions have been observed in practically all the quartz of the granodiorite. The euhedral quartz crystals in the quartz-alunite veinlets also contain many cavities with small well-defined cubes.

Microscopic description quartz-alunite veinlet.—Figure 4 sketches a quartz-alunite veinlet with visible flakes of molybdenite. The veinlet is 1.5 mm thick and cuts across an intergrowth of quartz and orthoclase, with a few brown hydromicas.

The quartz in the vein has apparently grown outward from the walls of the fissure. Some basal, hexagonal sections are 0.8 mm across. Alunite occurs both as a fine-grained aggregate of small, pseudo-cubic crystals ranging from .006 mm to 0.016 mm and as an aggregate of intimately intergrown, irregular grains. Birefringence colors are first-order grays and whites. Alunite fills the spaces between subhedral quartz crystals. While some contacts between quartz and alunite are straight with no sign of replacement, most show that quartz is corroded and irregularly replaced by alunite. The euhedral shapes of quartz, showing well-developed prisms with double rhombohedron terminations, suggest that the quartz grew unimpeded in open fractures, with alunite forming after the crystallization of quartz.

Molybdenite flakes line the sides of the quartz-alunite veinlet, though a few crystals may be in the center of the vein. Most individual molybdenite crystals are rectangular with lengths of 0.03 mm–0.06 mm and widths of 0.006 mm–0.03 mm. Most flakes lie with their long dimension parallel to the walls of the vein. Nearly all molybdenite occurs in alunite, but few flakes may be found in quartz, and yet fewer in the wall rock. Perhaps the most noteworthy feature is that most molybdenite is concentrated along the margins of veinlets. Writing about the molybdenite deposit at Climax, Colorado, Staples and Cook (1931, p. 13) state, "If we study any clear cut quartz veinlet we find the molybdenite lining the two sides of the veinlet becoming very scarce in the center." To account for the small size of the MoS_2 flakes, as well as their marginal occurrence in quartz veinlets, these authors suggest that molybdenite was introduced in colloidal form along with quartz.

At Santa Rita it seems clear that the molybdenite did not form with the quartz although it may also be colloidal. Figure 4 shows euhedral quartz in discrete units rather than in fine-grained, interlocking texture due to crystallization of a gel. It also reveals the tendency for molybdenite to occur in alunite rather than quartz. This suggests that MoS_2 was introduced with alunite after the formation of quartz. One can only speculate on the reasons for the concentration of molybdenite along the margins of the veinlets. Probably migration of alunite and molybdenite was principally from the center of the vein outward, the cooler walls of the veins causing precipitation of the molybdenite when the concentration of molybdenum sulfide became sufficiently high.

Microscopic description quartz-sericite-pyrite veinlet.—The quartz-sericite-pyrite veinlet pictured in Figure 5 is approximately 3 mm thick. It cuts across granodiorite containing hydromica phenocrysts. Feldspars altered to kaolinite are also porphyritic in a groundmass of quartz and orthoclase. Near the veinlet the kaolinite has been converted to sericite. Thin sections indicate that the veinlets cut across altered feldspar phenocrysts.

Quartz occurs as euhedral crystals, in places interlocking, and averaging 0.1 mm diameter. This is also the average size of the quartz in the groundmass of the host. The quartz is irregularly distributed in the vein but tends to be associated with pyrite. Sericite forms a border zone along each wall of the vein and also occurs interstitially between quartz grains. Sericite flakes are also found in the groundmass and as partial replacement of clay minerals in altered feldspars of the granodiorite.

In Figure 5, the pyrite is shown as two thin, discontinuous bands within the vein. More commonly the sulfide is a single, central seam in the quartz-sericite veinlet. Most pyrite shows replacement by chalcocite along grain boundaries or as infiltrations along cracks.

Structural Trends of Veinlets

Exposures containing the network of veinlets indicate that the veinlets in the altered rock are aligned preferentially along definite structural trends. No statistical treatment of their occurrence and distribution was attempted for this would be a major task, but enough strikes and dips of such veinlets were measured to indicate the major structural directions. These measurements indicate two principal trends and two subordinate zones (Table 2). Table 2 suggests that the barren quartz veins follow two main structural planes. Most quartz-sericite-pyrite veins favor the N.65°E. strike with 30°NW. dips and the N.20°W. strike with 80°SW. dip, while a few chalcocite veinlets follow a. N40°W. trend. From vantage points high above the south pit, and at different times of day, these dominant structural trends become strikingly apparent through light and shadow effects on many parallel rock surfaces. A set of vertical joints trends N.65°E. No veinlets were found along this direction.

Relative Age of Veinlets

The three types of veinlets described bear a consistent structural relationship to each other. The barren quartz veins are the least continuous—*i.e.*, they are displaced by other two types. Outcrops show that quartz-alunite and quartz-sericite-pyrite veins displace the barren quartz veins, but the reverse is never observed. Field study also shows that quartz-sericite-pyrite veins consistently cut the quartz-alunite type. Thus the barren quartz veins were the earliest, the quartz-alunite veinlets followed, and the quartz-sericite-pyrite type was the last to develop.

The early barren quartz veinlets cut across stage 2 granodiorite areas in the south and north pits. They are apparently not so common in the least altered part of the intrusive exposed on the north benches of the north pit. Here aplite dikes and nearly horizontal sheets traverse the fresh-looking granodiorite. Kerr *et al.* (1950, p. 294) noted that quartz veins intersect both aplite and fresh granodiorite porphyry. Seemingly the quartz veins formed soon after the main crystallization period of the intrusive and the injection of the aplite.

TABLE 2.—STRUCTURAL TRENDS OF VEINLETS

Quartz Quartz-sericite-pyrite	Quartz Quartz-alunite Quartz-sericite-pyrite	Chalcocite veinlets	Jointing

Microscopic Description of Hornblende Dike

On the east wall of the lowest bench in the south pit (5858 Bench at the time of study), a strongly bleached dike, about 4 feet wide, cuts across stage 4 porphyry. Hand specimens of the dike are bleached white by alteration. Pseudomorphous, white, fine-grained material after feldspars, rounded quartz, and light-brown phenocrysts indicate porphyritic texture. Dispersed in the dike are quartz fragments half an inch to an inch long and approximately half an inch in width. Most are rounded, though a few

are subangular. Some larger inclusions resemble fragments of stage 4 granodiorite containing quartz veinlets.

The bleached dike is characterized in thin section by altered biotite, hornblende, and feldspar phenocrysts in a fine-grained groundmass. Absence of quartz phenocrysts is characteristic.

The former presence of hornblende is deduced from the occurrence of typical hornblende cross sections, as well as narrow and elongated blades given by longitudinal sections. The latter have a length-width ratio of 3:1 (Fig. 6). Another bladelike mineral with a maximum length to width ratio of 2:1 is thought to represent altered biotite. The feldspars look fresh in ordinary light but under crossed nicols are completely altered to clay and sericite. Quartz phenocrysts are characteristic of the altered granodiorite porphyry but are lacking in the dike. This suggests that the unaltered dike was petrographically different from the granodiorite intrusive.

Significance of Hornblende Dike

The lack of quartz phenocrysts and the relative prominence of altered hornblende in the bleached dike agree well with the description of hornblende granodiorite porphyry dikes near Hanover, New Mexico (Kerr *et al.*, 1950, p. 294–298). These authors suggest that the hornblende porphyry dikes are earlier than the Hanover intrusive. At Santa Rita, structure indicates that the bleached, hornblende-bearing dikes are later than the granodiorite intrusive. The quartz-rich granodiorite dikes vary in age between Hanover and the Ground Hog mine, and the same may be true of the hornblende dikes.

While the altered dike cannot be unequivocally identified with the hornblende granodiorite porphyry dikes of the region, the petrographic similarity strongly suggests their equivalence. If the bleached dike in the Santa Rita pit does represent a hornblende porphyry dike, it would provide important structural and mineralogical evidence. The intrusive character of the dike indicates that both hornblende- and quartz granodiorite porphyry dikes are later than the Santa Rita granodiorite intrusion. The stronger alteration of the hornblende dike compared to that of the quartz porphyry dikes suggests that the former is the older. Structural evidence (Kerr *et al.*, 1950, p. 299) confirms this sequence.

The hornblende granodiorite porphyry dikes near Hanover are closely related to zinc mineralization of adjacent, favored limestone beds. Where zinc deposits are associated with the hornblende dikes, the latter have been altered to epidote, chlorite, sericite, and calcite (Kerr *et al.*, 1950, p. 332–335). No such alteration minerals were found in thin sections of the hornblende dike from the Santa Rita pit. It may be assumed that they formed as early alteration products related to zinc mineralization and were later obliterated by hydrothermal solutions associated with copper mineralization. Yet it seems as logical to believe that epidote and the other minerals never formed, and that the dike has undergone only hydrothermal alteration related to copper mineralization.

Added support for the latter view is the zonal distribution of copper and zinc around the Santa Rita intrusive. Figure 7, a geological map of an area north of the Santa Rita pit, shows the location of diamond-drill holes and the distribution of copper and zinc based on drill-hole data. A broad-scale zonal distribution of copper and zinc is suggested, although in detail boundaries between copper, copper-zinc, and zinc areas fluctuate toward and away from the igneous intrusive owing to the combined effect of hornblende dike occurrence, presence of premineralization fractures, and variations in proximity to the igneous contact in depth. The same pattern apparently occurs west of the pit, but more data are needed to verify its existence elsewhere around the Santa Rita granodiorite intrusion. The concept of zonal distribution of copper and zinc seems applicable to the Santa Rita intrusive. On this basis zinc mineralization and its related type of rock alteration would not be expected within the Santa Rita granodiorite. Therefore, the hornblende dike in the lower bench of the south pit was probably never subjected to epidotization, and shows only the effects of hydrothermal alteration which accompanied copper mineralization.

Probably other hornblende granodiorite dikes may be found in the Santa Rita area, and float specimens from other locations in the

pit suggest this, but their extreme alteration makes them difficult to distinguish from the granodiorite porphyry, unless they are unusually well exposed by recent mining activity.

spread hydrothermal alteration effects of the porphyry. The dike rock is bleached, ferromagnesian minerals have been altered to light-brown hydromica pseudomorphs, and the

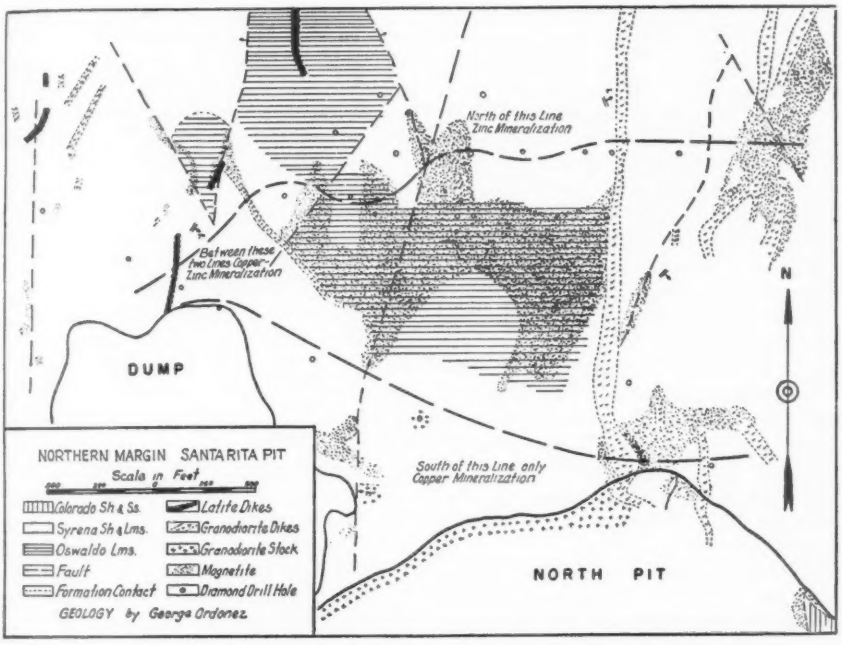

FIGURE 7.—COPPER-ZINC ZONING NORTH OF THE SANTA RITA PIT

Significance of Veinlet Sequence

The hornblende dike is free from barren quartz veins, although the intruded stage 4 porphyry is intricately veined by thin quartz seams. This lack of barren quartz veinlets, together with its bleached appearance, distinguishes the dike from the adjacent porphyry. However, quartz-alunite and quartz-sericite-pyrite veinlets cut both granodiorite and dike rock, suggesting that fracturing of the intrusive and associated quartz-vein filling happened early in the history of the deposit, and that the quartz vein episode was separated in time from the formation of the metal-bearing veins. This is further evidence for the relatively early age of the barren quartz veins.

Although the dike lacks the early barren quartz veinlets, it is not free from the wide-

feldspars are converted to clay. Since both dike and adjacent porphyry have been hydrothermally altered, the pervasive rock alteration of the granodiorite intrusive presumably took place after the intrusion of the dike and hence after the main period crystallization and formation of the early barren quartz veinlets.

The sequence of veinlets suggests another conclusion. As both quartz-alunite and quartz-sericite-pyrite veinlets cut the highly bleached dike and stage 4 granodiorite, metallic mineralization of the veinlet type probably took place after the main phase of hydrothermal alteration of the igneous rocks. It is also possible, however, that the later veinlets are contemporaneous with the alteration. Pyrite veinlets cut all stages of altered rock, even the least-altered phase of stage 1 granodiorite porphyry.

Hydrothermal Alteration of the Santa Rita Intrusive

Mineralogic Changes

Previous mineralogical studies.—The mineralogy of the hydrothermal alteration of the Santa Rita porphyry has been treated in detail (Kerr, Kulp, Patterson, and Wright, 1950). Four mappable alteration units, ranging from nearly fresh to almost completely destroyed rock, were arbitrarily selected in the field, but subsequent laboratory studies enabled correlation of stages 1 to 4 with definite forms of wall-rock alteration. Stage 1 represents the least-altered phase of the granodiorite porphyry, in which biotite phenocrysts are black and lustrous, and feldspar phenocrysts give a vitreous reflection. Stages 2 and 3 belong to the argillic type of alteration, where the feldspars are changed to clay minerals, biotite alters to chlorite and subsequently to hydromica, while the groundmass is argillized. Stage 4 corresponds to the development of quartz and sericite. Relict feldspars and porphyritic hydromica pseudomorphs are obliterated, and the groundmass becomes sericitized and silicified. The end-product of alteration is apparently a highly silicified rock.

Additional features.—Examination of about 80 thin sections brought out additional features. Slides of the least-altered granodiorite showed that even this fresh-looking rock is slightly altered. Most biotite phenocrysts are partially altered to chlorite (Fig. 8).

In some stage 2 material, two generations of biotite are indicated (Fig. 9). The small laths of secondary biotite are considered alteration mica, formed at the expense of earlier chlorite. Peterson *et al.* (1946, p. 834) note similar recrystallization biotite in the Castle Dome deposit, Arizona. More recently, Sales and Meyer (1948, p. 19) mentioned it in the kaolinite subzone of wall-rock alteration at Butte, Montana. They note that the chlorite fringe around unaltered residual biotite was in turn changed to an alteration biotite.

A striking texture common in hydromica pseudomorphs of stage 3 alteration consists of a grid of delicate rutile needles within altered mica phenocrysts (Fig. 10). The rutile needles are 0.04 mm–0.08 mm in length and only 0.004 mm in width. They are oriented parallel to three directions at angles of 60°. A similar grid in star mica has been attributed to the parallel growth of a multitude of tiny rutile or tourmaline needles on the 001 plane of mica (Buckley, 1951, p. 410). The formation of rutile in the alteration of biotite has been attributed by Sales and Meyer (1948, p. 16) to displacement of titanium from biotite by base exchange. At Butte separation of titanium dioxide occurred in the montmorillonite subzone of alteration, but the rutile grid at Santa Rita was observed only where kaolinite is the dominant clay mineral. Possibly some titanium dioxide was also derived from the breakdown of accessory sphene, which is present in the least-altered phase of the granodiorite but disappears as alteration increases. In stage 4 granodiorite, irregular concentrations of stubby, semi-rounded, reddish rutile crystals are thought to represent former rutile grids which have been rearranged, dispersed, and possibly corroded during the quartz sericitization of the granodiorite.

Interpretation of differential thermal analyses.—Differential thermal analyses of altered granodiorite assayed for copper have been sought for a good picture of the mineralogy of progressive alteration. However, it is difficult to separate clay and other constituents. The hydrothermally altered rock remains coherent; only stage 2 specimens permit selective hand picking of material. Fine grinding, elutriation, and heavy-liquid treatment effected only partial separation of alteration minerals, and the resulting mixture could not be interpreted decisively. Through painstaking efforts some altered material from three samples was isolated sufficiently to give recognizable peaks (Fig. 11). Additional time-consuming separations did not seem warranted. Thermal curves revealed approximately equal amounts of montmorillonite and illite in stage 2 samples. Kaolinite is the dominant clay mineral in stage 3. This curve also shows an exothermic peak between 375° and 500° C., owing to finely disseminated sulfides. The sample of the quartz-sericite type of alteration shows a prominent sulfide exothermic peak between 400° and 550° together with a strong endothermic quartz peak at 575°. Clay minerals were evidently absent.

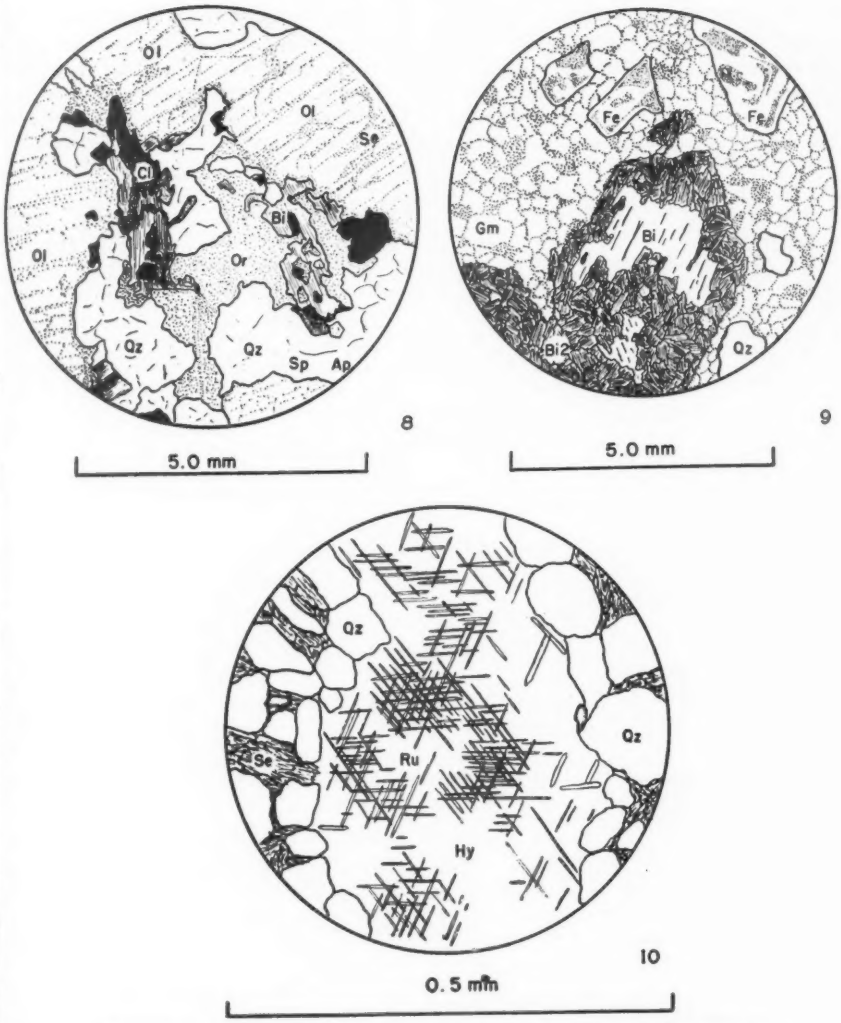

FIGURE 8.—CHLORITIZATION OF BIOTITE—STAGE 1, GRANODIORITE PORPHYRY

Oligoclase (Ol), orthoclase (Or), quartz (Qz), biotite (Bi), chlorite (Cl), sericite (Se), sphene (Sp), apatite (Ap), magnetite black.

FIGURE 9.—SECONDARY BIOTITE—STAGE 2 GRANDIORITE PORPHYRY

Feldspars altered to montmorillonite-illite (Fe), quartz (Qz), original biotite (Bi), secondary biotite (Bi2), groundmass of quartz and orthoclase (Gm).

FIGURE 10.—RUTILE GRID IN HYDROMICA—STAGE 3 GRANODIORITE PORPHYRY

Hydromica-illite (Hy), quartz (Qz), sericite (Se), rutile (Ru).

Chemical Changes

Four assay splits, derived from four bulk samples in separate areas of altered granodiorite, were analyzed chemically (Table 3) to determine chemical changes produced in wall rock by hydrothermal alteration. The analyses

are not representative of chemical reactions involved in the four alteration stages but should reflect quantitatively the gains and losses in the granodiorite porphyry during progressive alteration. Complete chemical analyses were omitted since other studies (Sales and Meyer, 1948, p. 16) indicate that the determination of a restricted number of oxides and elements is sufficient to show significant changes of wall-rock alteration. SiO_2, Al_2O_3, FeO, MgO, CaO, K_2O, Na_2O, H_2O, CO_2, S, and Cu only were determined.

Weight percentages of oxides do not give a true picture of chemical gains and losses involved in hydrothermal alteration of the granodiorite porphyry. The composition of the altered rock is more accurately expressed in grams per cc, assuming no changes in volume between fresh and altered rock (Sales and Meyer, 1948, p. 16). In Figure 12 the composition of the altered granodiorite has been recalculated as grams per cc. The changes involved in wall-rock alteration are indicated by a line connecting weight in grams of each radical in the four samples. No reference line giving average composition of fresh granodiorite is shown because (1) no exposures of unaltered granodiorite were available, and (2) not many chemical analyses were made. It is thought that the composition of the stage 1 sample approximates that of fresh granodiorite, because alteration of specimens from location A-5 is slight.

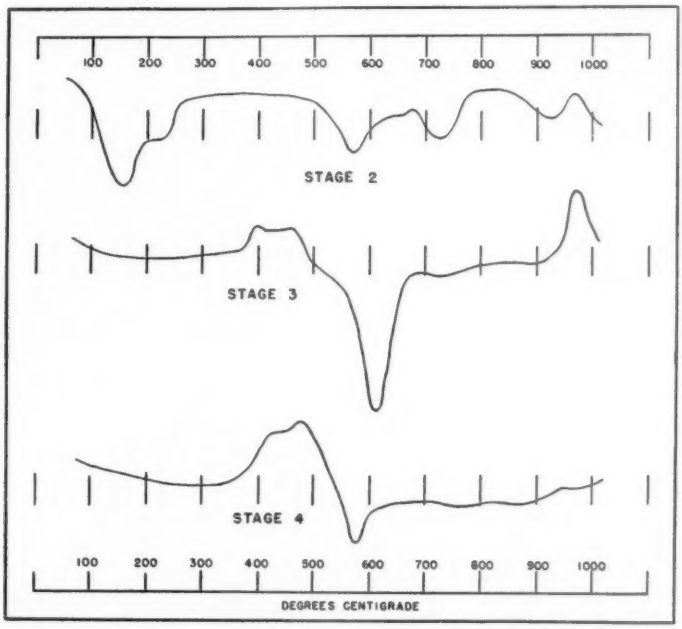

FIGURE 11.—THERMAL CURVES OF ALTERED GRANODIORITE PORPHYRY

Figure 13 shows that the specific gravity of the granodiorite decreases considerably in stages 2 and 4. In stage 2 the decrease is apparently due to development of clay of expanding lattice type, while in stage 4 it is probably the result of strong sericitization. The decrease of silica in sample E-3 as compared to stage 3 sample F-1 confirms the microscopic observation that sericite is better developed than silica in the stage 4 sample. In this sense E-3 does not represent the average stage 4 sample, which probably would indicate an increase of silica over the average stage 3

and less-altered granodiorite. Fixation of K_2O in the quartz-sericite stage of alteration is well shown by the rise of the potash curve. Alumina remains fairly constant, following a slight loss in stage 2. Ferric iron is approximately the same in the least- and most-altered granodiorite sample. In stage 2 it decreases to below the value of stage 1, but it increases again in stage 3. Lime, soda, and magnesia decrease from stage 1 through stage 4 and reflect leaching of these constituents by argillization and sericitization. Ferrous oxide and carbon dioxide follow a similar trend. The development of montmorillonite-illite types of clay is indicated in the rise of the H_2O curve for the stage 2 sample. Sulfur and copper curves rise steadily through argillization, but then level off in sericitization. The high level of the copper curve in the stage 4 sample should not be interpreted as strong hypogene copper mineralization in this rock.

Summarizing, in the argillic type of alteration there is a pronounced loss of MgO, CaO, and Na_2O with an associated decrease in silica. Relative gains in this zone are registered in water of hydration and sulfur. In the sericite zone, leaching of MgO, CaO, and Na_2O continues, while potash, sulfur, and probably silica increase. Chemical changes involved in progressive hydrothermal alteration of the Santa Rita granodiorite porphyry intrusive are remarkably similar to those established at Butte, Montana (Sales and Meyer, 1948; 1950).

Copper Mineralization of the Santa Rita Intrusive

Types of Copper Mineralization

General discussion.—Copper mineralization is of several types at Santa Rita, though probably only one is of economic importance. The metal is recovered principally from chalcocite in altered sediments and sills surrounding the granodiorite intrusive and also within the intrusive itself. The bulk of the ore is derived from the latter. Basic copper carbonates, cuprite, and native copper are still found today, but they are of mineralogic rather than economic interest. Sheets of native copper several inches long and half an inch thick were found in the lowest bench of the south pit along fractures through which acidic water was flowing. This indicates that fissures and other structural conduits may locally carry surface oxidation to depths of several hundred feet. Excluding copper minerals of the oxidation zone, the remaining forms of copper mineralization may be arbitrarily classified:

(1) quartz-sericite-pyrite veinlets
(2) Pyrite-chalcocite veins
(3) disseminated pyrite-chalcocite
(4) disseminated chalcopyrite
(5) disseminated chalcopyrite-bornite

Quartz-sericite-pyrite veinlets.—Chalcocite has been precipitated on pyrite and has also replaced it to some extent. Polished surfaces of specimens from this type of veinlet were examined to determine the nature of this form of copper mineralization.

The pyrite seam in the center of the veinlet encloses euhedral quartz crystals. Pyrite does

Table 3.—Chemical Analyses of Granodiorite Porphyry

(Silve Kallmann, Analyst)

	Stage 1 A-5	Stage 2 B-1	Stage 3 F-1	Stage 4 E-3
SiO_2	64.87	64.58	70.14	68.75
Al_2O_3	16.63	17.82	14.86	15.38
Fe_2O_3	2.36	1.16	2.36	2.03
FeO	2.35	1.14	0.66	0.57
MgO	2.27	1.93	1.08	1.16
CaO	3.24	0.93	0.05	0.05
Na_2O	5.71	0.76	0.14	0.14
K_2O	2.79	3.11	2.77	3.97
H_2O	0.15	0.74	0.15	0.10
CO_2	0.08	0.04	0.05	0.03
S	0.22	0.62	1.86	1.35
Cu	0.07	0.15	0.71	0.92
Total	100.74	93.98	94.83	94.45
Specific gravity	2.36	1.94	2.41	2.20

Designations A-5, B-1, F-1, E-3 refer to sample and specimen locations as shown on Figure 15.
(A-5) Hard, unaltered rock from north pit.
(B-1) Soft, friable sample from 5858 Bench south pit.
(F-1) Bleached rock near quartz granodiorite Dike 5858 Bench, south pit, west wall.
(E-3) Highly altered rock from 5984 Bench, south pit.

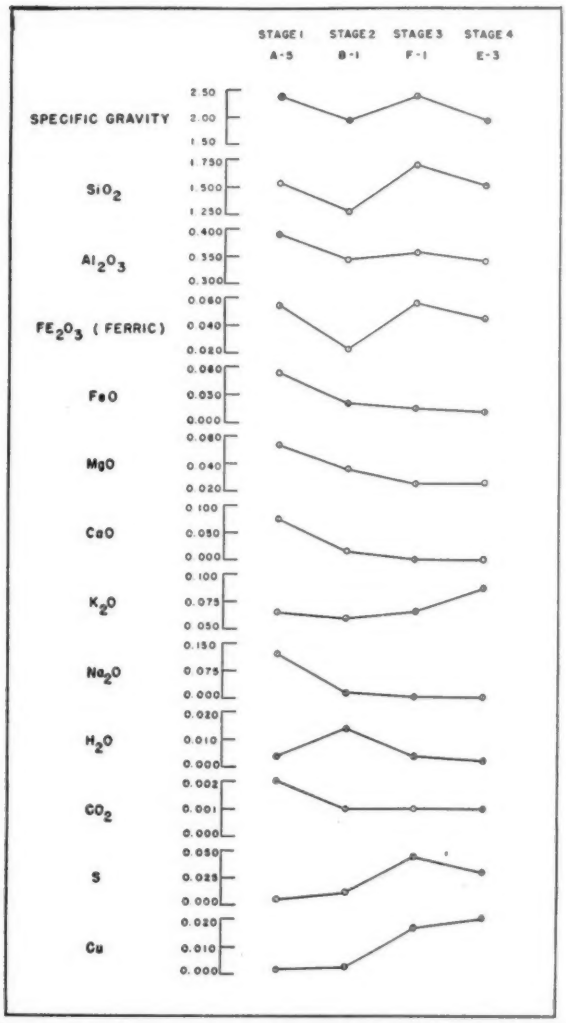

FIGURE 12.—CHEMICAL COMPOSITION OF ALTERED GRANODIORITE PORPHYRY EXPRESSED IN GRAMS PER CUBIC CENTIMETER

not form a continuous stringer of uniform thickness, but pyrite masses are separated by quartz areas at irregular intervals. Apparently pyrite was deposited in voids of the vein, left after crystallization of quartz. The quartz, pyrite, and sericite, however, are thought to have formed simultaneously as suggested by Sales and Meyer (1950, p. 263) in the sericite zone of alteration at Butte, Montana. Pyrite also forms small, isolated grains within the halo of sericite alteration. Both types of pyrite contain many minute chalcopyrite inclusions (Pl. 1, fig. 1), some of which are recognized only at high magnifications. The inclusions measure about 35 microns across. Most chalcopyrite inclusions are elliptical and have rounded, smooth contacts with pyrite. In places, some pyrite contains similarly shaped inclusions of

chalcocite. These are thought to represent selective replacement of chalcopyrite by chalcocite. Figure 2 of Plate 1 is a microphotograph of a cross sections of vermicular chalcopyrite veinlets cutting pyrite, but lack of such veinlets in all polished sections examined argues against

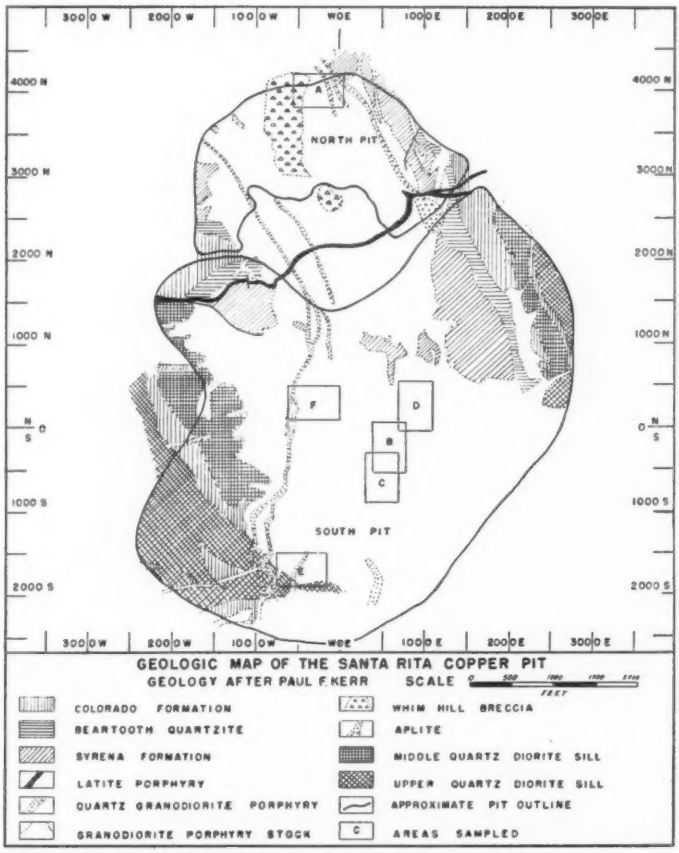

FIGURE 13.—GEOLOGICAL MAP OF SANTA RITA COPPER PIT

chalocpyrite inclusion magnified 1500 times, revealing the selective replacement of the copper-iron sulfides.

The origin of the inclusions is not entirely clear. Possibly they represent unreplaced remnants of chalcopyrite in pyrite. That pyrite is the replacing mineral is supported by the presericite age of hypogene chalcopyrite. Some disseminated primary copper sulfides may have been partly replaced by pyrite, formed during late sericitization of parts of the intrusive. The rounded inclusions might be only this. Furthermore, if chalcopyrite veinlets were present, they should be the first to be replaced by chalcocite, yet most inclusions are free from chalcocite replacement. Wherever chalcopyrite blebs show partial or complete chalcocitization, there are minute fractures that permit access to enriching solutions. The elliptical or rounded shape of the inclusions, their smooth contact with pyrite, and their irregular distribution suggest unmixing from solid solutions. Edwards (1947, p. 35–93) discusses solid solutions in ore minerals but

does not mention chalcopyrite-pyrite. Solid solutions of chalcopyrite and pyrite should form with difficulty since these two compounds do not meet the requirements for isomorphism (Rankama and Sahama, 1950, p. 121). They do not have similar atomic structures, and their cation-anion ratios differ.

Pending special study of the behavior of pyrite-chalcopyrite intergrowths at elevated temperatures after the manner of Schwartz (1927), it is unlikely that the observed texture indicates unmixing from a solid solution.

Segregation of chalcopyrite in pyrite as a result of postulated immiscibility of the two sulfides is not likely in view of the probable origin of the sulfides from hydrothermal solutions or gaseous emanations.

Some polished surfaces show that minor molybdenite also occurs in the veinlets. The small, bladed outlines of these crystals occur on the edges of the medial pyrite vein. The occurrence of molybdenite indicates a relatively high temperature of formation of minerals directly related to molybdenite. Its presence in decreased amounts, compared with the earlier quartz-alunite veinlets, suggest that the temperature of the mineralizing solutions had decreased by the time most quartz-sericite-pyrite veins formed, and that molybdenite deposition gave way to the formation of pyrite.

The main copper content of the veinlets is not derived from the small inclusions of chalcopyrite, but from chalcocite which coats or partially replaces pyrite. Figure 3 of Plate 1 shows the margin of a pyrite stringer, with nearly euhedral quartz in the pyrite. Chalcocite replacement is better developed on the margins of the pyrite seam at the left of the microphotograph than toward the center, indicating that replacing solutions worked their way into the veinlet from the outside. Chalcocite replacement has apparently been facilitated along boundaries between quartz crystals and pyrite and, on a larger scale, between the pyrite veinlet and the sericite envelope. Some polished surfaces show almost complete replacement of pyrite by chalcocite. When etched with KCN, chalcocite shows a very fine-grained, granular texture in which some grains have developed parting in one direction, characteristic of orthorhombic chalcocite.

Pyrite-chalcocite veinlets.—A group of thin veins which cut the altered rock in one of the lower benches of the south pit lack the conspicuous alteration halo of the quartz-sericite-pyrite type of veinlets and show knife-sharp contacts with country rock. This type of veinlet, up to 1 inch in width, consists of the "steely" variety of chalcocite and pyrite. The massive chalcocite contrasts greatly with the dull coating or thin replacement of pyrite in the quartz-sericite veinlets. Some veins contain only chalcocite, and individual veins bifurcate upward and pinch out along one branch.

Microscopic study of polished surfaces reveals classic examples of chalcocite replacing

PLATE 1.—PHOTOMICROGRAPHS OF POLISHED SURFACES

FIGURE 1.—PYRITE GRAIN WITH CHALCOPYRITE INCLUSIONS

Two inclusions appear to have been replaced by chalcocite. Chalcocite rims pyrite. Dark-gray areas are gangue. Plain reflected light, ×100.

FIGURE 2.—SELECTIVE REPLACEMENT OF CHALCOPYRITE INCLUSION IN PYRITE

Chalcocite darkest. Plain reflected light, ×1500.

FIGURE 3.—CHALCOCITE REPLACED BY PYRITE IN QUARTZ-SERICITE-PYRITE VEINLET

Dark-gray areas are quartz, lighter gray chalcocite, and nearly white pyrite. Plain reflected light, ×100.

FIGURE 4.—PYRITE REPLACED BY CHALCOCITE IN PYRITE-CHALCOCITE VEIN

Plain reflected light, ×700.

FIGURE 5.—KCN ETCH PATTERN ON CHALCOCITE FROM PYRITE-CHALCOCITE VEIN

Etched chalcocite shows orthorhombic pattern. Plain reflected light, ×700.

FIGURE 6.—CHALCOPYRITE AND COVELLITE IN SILICEOUS STAGE 2 GRANODIORITE PORPHYRY

The dark-gray blades are covellite in chalcopyrite. The smooth, light-gray mineral is pyrite. Plain reflected light, ×200.

FIGURE 1

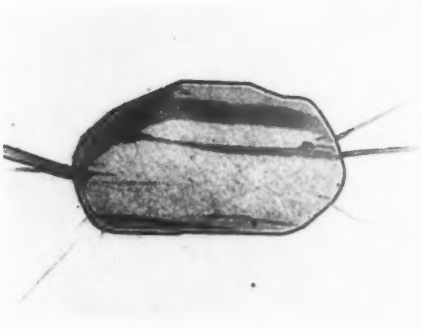

FIGURE 2

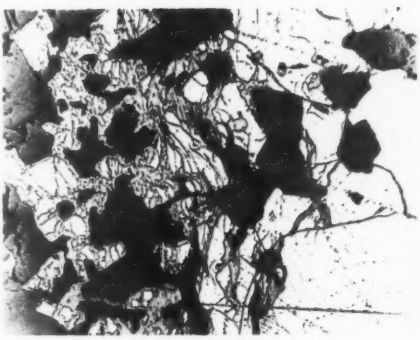

FIGURE 3

FIGURE 4

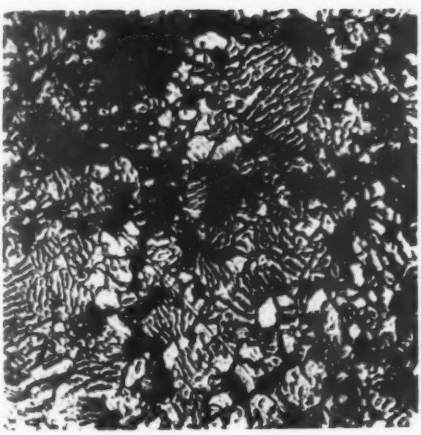

FIGURE 5

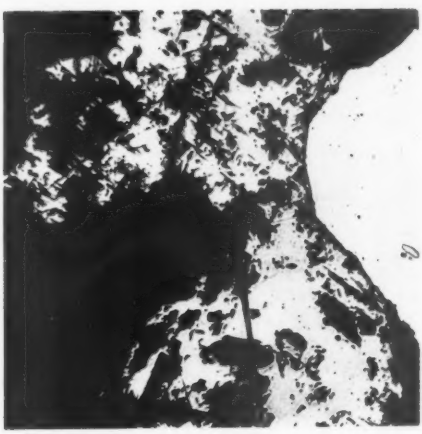

FIGURE 6

PHOTOMICROGRAPHS OF POLISHED SURFACES

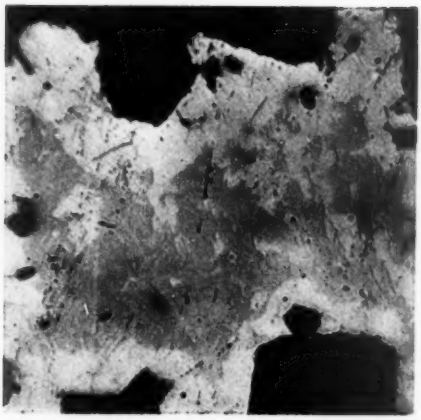

FIGURE 1

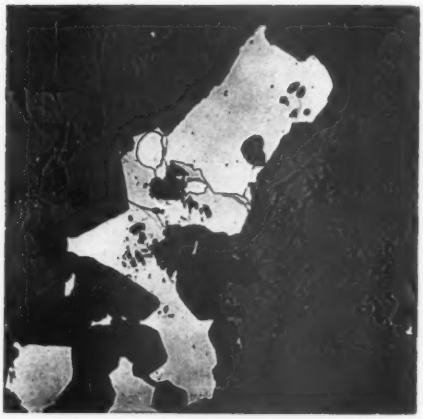

FIGURE 2

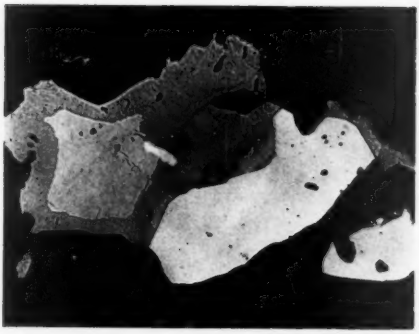

FIGURE 3

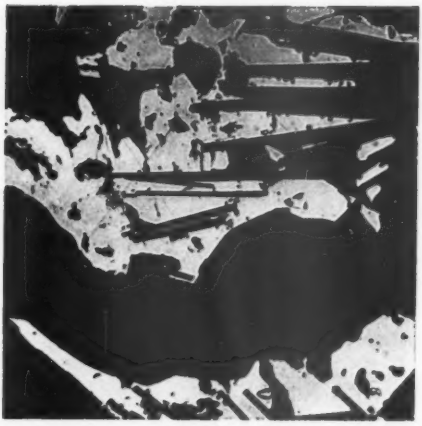

FIGURE 4

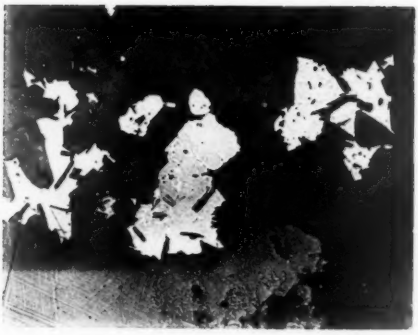

FIGURE 5

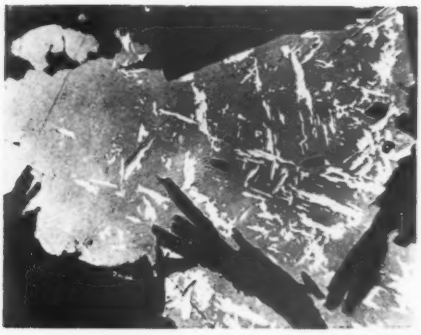

FIGURE 6

PHOTOMICROGRAPHS OF POLISHED SURFACES

pyrite (Pl. 1, fig. 4). Replacement extends across the width of the vein, suggesting that the vein once was solid pyrite. All degrees of replacement may be observed, and in some specimens only minute specks of pyrite remain. In all sections examined the chalcocite is uniformly gray, and no covellite was detected. To verify the nature of the chalcocite, some surfaces were treated with KCN. A granular texture, somewhat coarser than that of chalcocite from quartz-sericite-pyrite veinlets, developed on the polished surfaces. Most etched-out grains show parallel lines, the pattern of orthorhombic chalcocite (Pl. 1, fig. 5). X-ray spectrometer study, though not conclusive, also indicates orthorhombic chalcocite. A few lines of low intensity could not be related to the orthorhombic pattern and may represent a different phase of chalcocite. Possibly, also, these lines reflect a structural change resulting from fine grinding.

The pyrite-chalcocite veins occur in an area of stage 4 alteration where visible quartz has been introduced, but where no quartz-sericite-pyrite veinlets penetrated, so that their mutual relationship could not be established. The chalcocite veins definitely cut stage 4 rock and are therefore later than the widespread hydrothermal alteration. The rhombic chalcocite is thought to be hypogene because (1) chalcocite is massive and steely, (2) chalcocite replaces pyrite through the width of the vein, not just along its margin, (3) veins of this material bifurcate and pinch out upward. The orthorhombic phase of chalcocite persists to approximately 105° C. (N. Buerger, 1941, p. 39).

Disseminated pyrite-chalcocite.—Perhaps the simplest and most common form of copper mineralization is chalcocite, typically replacing the rims of isolated pyrite grains. All degrees of replacement may be found, and small disseminated grains of chalcocite without cores of pyrite are also numerous. Pyrite may contain minute chalcopyrite inclusions, then resembling Figure 1 of Plate 1, but many grains do not contain such inclusions, or at least no inclusions are visible with the reflecting microscope.

It is difficult to determine whether most pyrite-chalcocite grains are truly disseminated in the porphyry—*i.e.*, randomly distributed throughout the rock without any apparent guidance by fractures or other structural planes. Fractures in the deposit range from breaks of major magnitude to the veinlets described earlier, and even to smaller fissures which cause nearly all hand specimens to break into sharp, angular fragments. In the field one sees that pyrite-chalcocite grains are more abundant along exposed fracture planes. More than 30 years ago Ransome (1919, p. 158) remarked

> "On the whole, however, the commonly applied name 'disseminated ore' is likely to be a little misleading unless it is understood that the dissemination is due to the minute and dispersive character of the fissuring and veining much more than to the development of sulfides in isolated grains throughout the rock mass."

PLATE 2.—PHOTOMICROGRAPHS OF POLISHED SURFACES

FIGURE 1.—BORNITE REPLACED BY CHALCOCITE

Darker gray is bornite. Plain reflected light, ×700.

FIGURE 2.—CHALCOPYRITE IN PYRITE VEINLET CUTTING ALTERED SYRENA SEDIMENTS

Pitted, dark-gray areas are magnetite. Chalcopyrite replacing pyrite (northeast area of veinlet). Black areas in veinlet consist of jasper and quartz. Plain reflected light, ×100.

FIGURE 3.—CHALCOPYRITE PREFERENTIALLY REPLACED BY CHALCOCITE

Core of chalcopyrite is surrounded by gray chalcocite. Pyrite is light gray. Plain reflected light, ×300.

FIGURE 4.—CHALCOPYRITE CUT BY SERICITE AND QUARTZ

Black blades are sericite; rounded areas, quartz. Plain reflected light, ×250.

FIGURE 5.—BORNITE-CHALCOPYRITE INTERGROWTH

Chalcopyrite forms light-gray streaks. Plain reflected light, ×150.

FIGURE 6.—BORNITE-CHALCOPYRITE INTERGROWTH

Texture suggests unmixing from solid solution. Sericite flakes penetrate the intergrowth. Enlarged from Figure 5 of Plate 2. Plain reflected light, ×750.

Disseminated chalcopyrite.—The importance of the chalcopyrite to the genesis of the Santa Rita porphyry copper deposit has been intimated by Kerr and his coworkers (1950, p. 331). Its presence in the ore deposit has not been demonstrated in the literature. In this study, microscopic examination of several polished surfaces revealed disseminated chalcopyrite. Field work indicates that little or not megascopic chalocpyrite exists within the porphyry mass.[1] It is therefore not surprising that its presence in the intrusive remained undetected.

Polished-surface examination of specimens collected along the benches of the south pit at various elevations proved that chalcopyrite is present only as inclusions in pyrite, except in area C, where some chalcopyrite occurs as disseminated grains, commonly replacing pyrite (Pl. 1, fig. 6). As a rule this chalcopyrite is replaced by a network of covellite blades, which in some cases has preserved only a minute core of chalcopyrite. As the blades of covellite develop from margins to centers of the chalcopyrite grains, this form of enrichment is considered supergene. Chalcocite is conspicuously absent. Pyrite is partly replaced by chalcopyrite, possibly of two generations—an early, minor phase in the form of inclusions in pyrite, and a later, more important phase during which chalcopyrite formed at the expense of earlier pyrite.

Mr. G. J. Ballmer kindly supplied specimens for microscopic study from a diamond-drill hole, driven in the bottom of the south pit, and polished surfaces were prepared and studied. Each contained disseminated chalcopyrite. In the samples taken slightly below the 5858 level of the south pit, chalcopyrite is deposited on earlier pyrite grains and partially replaces them. The close association of the two sulfides suggests that pyrite crystals acted as loci for the deposition of chalcopyrite. Pyrite probably crystallized first from the mineralizing solutions. Chalcocite preferentially replaces chalcopyrite and may thus remove all indications of hypogene chalcopyrite (Pl. 2, fig. 3). At greater depths chalcopyrite becomes more common and free from secondary sulfides (Pl. 2, fig. 4), and chalcopyrite and pyrite are the only visible metallic sulfides.

Chalcopyrite in the Santa Rita porphyry copper deposit is of major importance, not only in consideration of the origin of the chalcocite ore body, but also in view of its implications on the depth of economic mineralization.

Disseminated bornite-chalcopyrite.—A polished surface of a specimen from area E shows disseminated grains of bornite and bornite with intergrown chalcopyrite. Most bornite occurs as irregular pink masses, partially replaced by chalcocite (Pl. 2, fig. 1). Pyrite is partly or completely surrounded by mixed bornite-chalcocite areas. Scattered grains of bornite contain elongated blades, spindles, and irregular aggregates of chalcopyrite (Pl. 2, fig. 5). At first, replacement of the dominant bornite by chalcopyrite was suspected, but closer examination of the unusual texture suggests a different mode of origin. Under high magnifications (Pl. 2, fig. 6) the elongated chalcopyrite masses are parallel to three principal directions, creating a fairly definite triangular pattern. Most intergrown chalcopyrite occurs as spindles—*i.e.*, in somewhat drawn-out masses thinning at both ends. As spindles cross or touch each other they contract rather than enlarge at intersections. Some chalcopyrite forms small rounded blebs or larger irregular aggregates. The chalcopyrite blades lie predominantly within the bornite itself and not on a rim. Such features are accepted textural criteria of segregation or unmixing from solid solutions (Schwartz, 1931b, p. 761). Solid solution of chalcopyrite-bornite was demonstrated experimentally in 1931 by Schwartz. He concluded "... deposits showing intergrowths of chalcopyrite in bornite (not replacements) were formed at or above 475° C." (1931a, p. 200). Schwartz also explained the comparative rarity of these intergrowths in nature as due to (1) the temperature of formation of the solid solution which is probably higher than that at which most sulfide deposits form, and (2) the rapidity with which diffusion or unmixing takes place. Such an intergrowth in a mineral deposit is strong evidence that the deposit formed above 475° C. and cooled rapidly. The intergrowth of bornite-chalcopyrite observed at Santa Rita resembles an

[1] Since the above was written, thin chalcopyrite veinlets have been found in drill cores from the island separating the north and south pits.

occurrence in Queensland (Edwards, 1947, Fig. 79) more than it does the laboratory results obtained by Schwartz. Spindles of chalcopyrite in bornite produced synthetically are more euhedral than those at Santa Rita. Possibly the original unmixing texture resulting from initial cooling was later modified by slight reheating and recooling, enabling the spindles to grow a bit more, mainly along their shorter direction, and allowing more chalcopyrite to segregate into small rounded blebs.

The two possibilities, unmixing or replacement, give rise not only to divergent conclusions but also to conclusions of contrasting significance. If replacement is inferred, then the texture is fairly common. It is, therefore, a poor indicator of the temperature of formation of the mineral deposit. If the intergrowth is the result of unmixing from a solid solution of bornite-chalcopyrite, the texture is an additional case of unusual, natural phenomena. More important, the intergrowth gives valuable evidence on the temperature of formation of the Santa Rita porphyry copper deposit. It is not known whether similar bornite-chalcopyrite intergrowths occur throughout the deposit and, therefore, that the temperature of formation of the intergrowth applies to the whole deposit, but the unmixing texture would probably be found at least throughout the stage 4 area from which this specimen was collected. We may therefore conclude that copper mineralization of a significant portion of the deposit occurred at 475° C. or higher.

Bornite-chalcopyite intergrowths have also been observed at San Manuel, Arizona (Schwartz, 1949). Schwartz states (1949, p. 267), "... recent investigation indicates that these are a result of overheating in drying the concentrates". The specimens from Santa Rita, however, were picked directly from the rock face and were not subjected to any heat treatment.

Correlation of Copper Mineralization with Hydrothermal Alteration

Sampling of Stages of Alteration

To elucidate the correlation between a certain stage of hydrothermal alteration and copper mineralization found at Santa Rita by Kerr and his associates (1950, p. 326–332), a closely controlled sampling program was undertaken during the summer of 1951.

Representative areas from the four stages of rock alteration were chosen on the basis of geologic occurrence, accessibility, and freedom from strong surface oxidation effects (Fig. 13, A–F). The best exposure of the least-altered granodiorite porphyry is along the north wall of the north pit. Although there has been no mining on these benches for many years, stage 1 rock shows only superficial weathering. Area A, located in the most accessible bench, yielded good samples from the least-altered section of the intrusive. Stage 2 granodiorite porphyry outcrops in both north and south pits, but as this rock is more altered than stage 1 and is not so resistant to weathering, the north pit exposures are poor. Better outcrops occur in the lower benches of the east wall of the south pit (areas B and C). Argillic alteration is widespread, and areas B and F in the south pit were picked on the basis of their recent exposure (during the summer of 1951) by mining operations. Stage 3 and 4 in many places intermingle intricately, and well-exposed areas of stage 4 are not common. A good zone of quartz-sericite alteration (area F) was chosen in the bend of the horseshoe made by the south pit, at the elevation of the 5984 bench.

In each area sample stations were located either by Brunton and tape measurements or by transit with the co-operation of the Chino Engineering Department. Sample stations were marked 10 to 15 feet apart within individual zones of alteration. The location of each station in the selected areas of hydrothermal alteration is shown by small, areal maps in Figure 15. Beneath each map of alteration and sample location, the copper content of every sample is given in the form of a bar graph. The copper determinations were made through the courtesy of the local assay department. Samples were cut with hammer and moil on each face, approximately 4 feet above the talus accumulation at the foot of the particular bank. The samples were weighed to determine the most representative size. At first, 5-pound samples were cut, but fluctuation in assays greater than expected led to an investigation of the variation between size of

sample and copper content. In three locations samples of four sizes (5, 10, 25, 50 pounds) were taken. The result of these three sets of assays is plotted in Figure 14. Each set of samples is indicated by a line connecting individual assays in the group. Figure 14 shows that samples

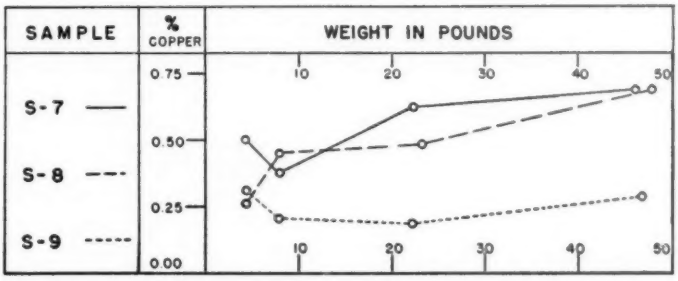

FIGURE 14.—VARIATION OF COPPER CONTENT WITH WEIGHT OF SAMPLES

of 25 pounds show no significant variations, while those under 25 pounds show great fluctuation; samples of approximately 25 pounds were therefore used. Discrepancies between assays in the smaller samples are due partly to incorporation or exclusion of mineralized veinlets. With larger samples, the effect of a mineralized veinlet is lessened.

Copper Content of Altered Granodiorite

The correlation between copper content and hydrothermal alteration of the granodiorite intrusive is shown graphically in Figure 15. The least-altered porphyry carries little copper. This is apparent even in the field, for the granodiorite is essentially a fresh-looking rock barren of metallic minerals except scattered magnetite and pyrite grains and some minute pyrite veinlets. In areas of stage 2 granodiorite porphyry hydrothermal alteration is more pronounced, and the rock is soft and friable. Good hand specimens are difficult to obtain, for the rock disintegrates on exposure. The outcrop in the south pit has a rounded outline in contrast to the angular fragments of the less-altered granodiorite. The character of the rock is linked directly to the nature of the minerals produced by hydrothermal alteration, in particular montmorillonite clay. Area B (Fig. 15) indicates the relative increase of copper in stage 2. Correspondingly one notices in the field an increase in the amount of pyrite in the rock.

Area C has been recorded as alteration stage 2, although the rock differs from that exposed in area B. The granodiorite exposures of area C were markedly more siliceous; the outline of the siliceous zone is readily visible because its peculiar, dark grayish-green color contrasts sharply with the brown and white of other altered areas. The darker color is apparently due to abundance of silica and pyrite and also to a lack of clay alteration products. The siliceous rock is also harder and tends to break along sharp, angular planes. Hand specimens reveal biotite and dull feldspar phenocrysts. Pyrite veinlets are numerous. Thin sections confirm the presence of small biotite laths and some chlorite, the partial alteration of the feldspars to sericite, and the abundance of silica both as veinlets and as irregular replacements of the whole rock. Except for the scarcity of clay minerals and unusually high silica contents, the alteration of the rock would be referred to stage 2, and the exposures in area C were so interpreted. The siliceous rock here is anomalous for stage 2 alteration and must represent an unusual phase. The siliceous zone carries considerably more pyrite than the other areas, yet pyrite is seldom coated with chalcocite, and consequently the copper content of this material is meager (Fig. 15). Pyritization appears to be later than the introduction of silica, for pyrite veinlets consistently cut quartz veins and quartz replacement areas. It was noted that barren quartz veinlets formed early after the consolidation of the intrusive, and this

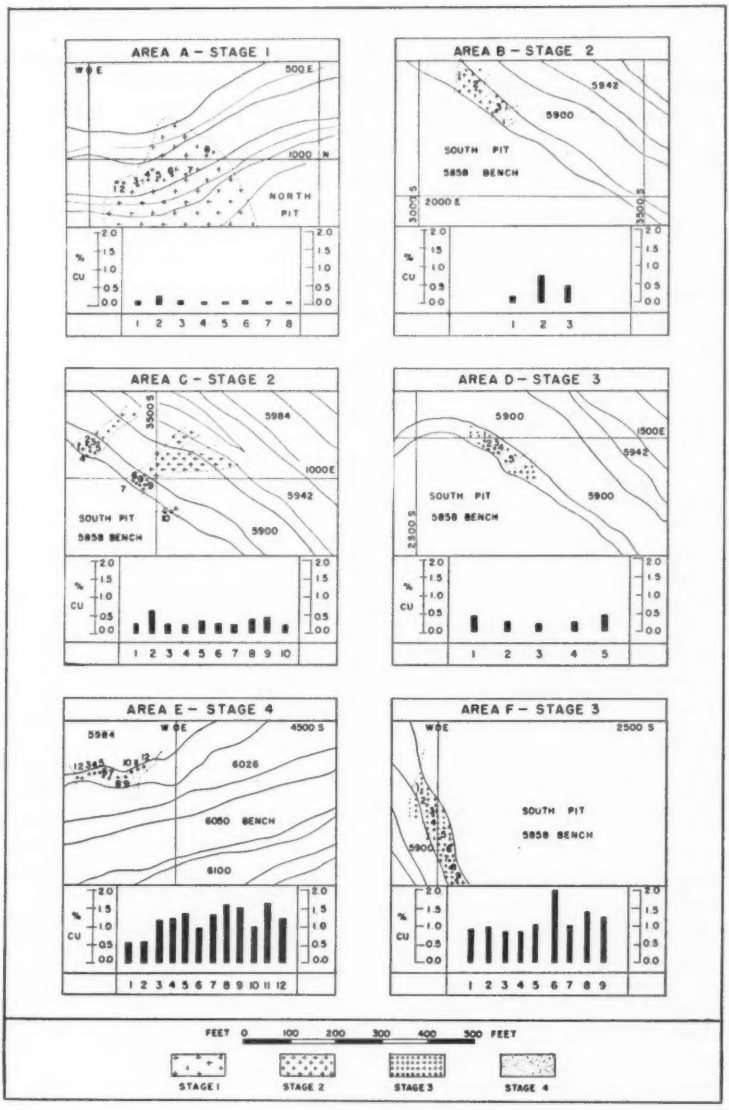

FIGURE 15.—COPPER CONTENT OF ALTERED GRANODIORITE

Numbers beneath columns of bar graphs refer to individual sample locations as shown on sketch maps of areas A–F. Copper content indicated by height of columns.

silicification may be related to the early quartz epoch. Possibly in area C introduction of silica was sufficient to yield not only quartz veins along definite channels, but also a partial silicification of the rock between the veins. Subsequently hydrothermal alteration became restricted to the unsilicified material, giving rise to a siliceous phase of stage 2 alteration.

The brittleness of this zone may account for its stronger fracturing, facilitating access and deposition of pyrite.

Kerr and his colleagues (1950, p. 327) first showed the concentration of copper values in stage 3 by superimposing their alteration map of an assay plat of stripping operations. Sampling of this stage of alteration shows a noticeable increase in the copper contents of some of this material (area F, Fig. 15) and a less marked rise in another area (area D, Fig. 15). Stage 3 granodiorite porphyry is a good ore bearer. The rock is bleached grayish white, differentiating it from its less-altered equivalents. Through much of stage 3 porphyry an intricate network of veinlets is found. Since some of these raise the tenor of the ore, the veinlets are of considerable importance and were studied in detail. With increased sericitization the rock begins to resemble stage 4, and through gradational steps develops into quartz-sericite type of altered granodiorite. With progressively stronger attack by hydrothermal solutions the hydromica pseudomorphs of stage 3 become rare and eventually disappear. When no hydromicas are observed the rock is recorded as stage 4. A good area was found on the 5984 bench (area F, Fig. 15). The rock is strongly bleached and is apparently not so silicified as other stage 4 areas. Thin-section observations indicate that the granodiorite has been strongly sericitized. Vague outlines of former feldspar phenocrysts may be recognized by clustering and parallelism of sericite flakes. Scattered quartz phenocrysts are the only original minerals of the granodiorite to escape intense sericitization and partial silicification. Assays of samples indicate the relatively high copper content of this stage of altered granodiorite porphyry. The high copper values seem to be due mainly to the disseminated pyrite coated with chalcocite and the presence of some bornite-chalcopyrite.

The sampling program confirms the occurrence of stronger mineralization within areas of more intense hydrothermal alteration. The program also furnished specimens for the laboratory investigation of the apparent correlation between copper mineralization and hydrothermal alteration.

Formation of Ore

In the relationship between copper mineralization and hydrothermal alteration in porphry copper deposits, it is of prime importance to determine whether material labelled "ore" is the result of hypogene metallization or of supergene enrichment (Schwartz, 1947, p. 351). At Santa Rita, this distinction is not easily established, as Kerr and associates (1950, p. 331–332) emphasized in their study.

Additional data on the ore-forming processes at Chino have been furnished by polished-section work on ore specimens from the deposit. It has been shown that chalcopyrite occurred in altered porphyry. This shows the existence of a possible copper source mineral for the chalcocite, which is the chief ore mineral in the deposit, and also cancels the need of postulating hypogene chalcocite as the main primary copper mineral. Evidence strongly suggests a supergene origin for most chalcocite and consequently for the ore body as well. All the chalcocite etched with KCN yields the low-temperature, orthorhombic form, though one minor chalcocite occurrence may be hypogene. The clearest indication that chalcocite is a secondary, downward-enrichment product is in the polished sections from a diamond-drill hole driven in the bottom of the pit. These specimens show that chalcocite disappears in depth with the gradual establishment of chalcopyrite as the dominant copper mineral. Similar evidence from other sections shows selective replacement of chalcopyrite and bornite by chalcocite. The copper source for chalcocite in the ore body may have been supplied by (1) disseminated chalcopyrite and bornite in the protore, (2) chalcopyrite inclusions in pyrite, and (3) chalcopyrite in pyrite veinlets cutting magnetite deposits in the sedimentary rocks in the contact zone of the intrusive (Pl. 2, fig. 7).

While chalcopyrite occurs in this manner in the peripheral parts of the intrusive, it is improbable that it formed above the porphyry mass, since the calcareous member of the Syrena formation was penetrated by the rising intrusive. No zone comparable to the limestone member of the Syrena occurs in the younger Beartooth or Colorado formations. It is there-

fore doubtful that there was an unusual concentration of chalcopyrite above the intrusion.

The total volume of chalcopyrite occurring as inclusions in pyrite is much less than that represented by the copper content of the primary copper sulfides. Disseminated bornite-chalcopyrite forms the principal source of copper for the development of supergene chalcocite. To produce the present ore body, the protore had to be enriched approximately fourfold.

Correlation of Hypogene Copper Mineralization with Hydrothermal Alteration

The secondary enrichment in the Santa Rita ore body obscures the correlation between hypogene copper mineralization and hydrothermal minerals. However, relics of chalcopyrite indicate the probable nature of this correlation. It has been observed that chalcopyrite becomes the dominant copper mineral in depth. Bornite, intergrown with chalcopyrite, also occurs in the mineralized porphyry so that we can postulate a low-grade, chalcopyrite-bornite hypogene mineralization for the Santa Rita deposit (with chalcopyrite more important quantitatively).

Some idea of the correlation of this primary form of copper mineralization with hydrothermal alteration of the granodiorite may be gained by examining Figures 4, 5, and 6 of Plate 2. Figure 4 of Plate 2 shows black, rectangular flakes of sericite projecting into the chalcopyrite. Similar sericite flakes cut chalcopyrite in other sections. Intergrowths of bornite and chalcopyrite (Pl. 2, fig. 5, 6) are also cut by sericite. In the more highly magnified picture the sericite crystals cut some exsolution spindles of chalcopyrite in bornite, indicating that sericite formed after the development of the sulfide intergrowth.

Geologic work at Castle Dome (Peterson *et al.*, 1946, p. 832) and at Silver Bell (Kerr, 1951, p. 479) substantiates the evidence from Santa Rita that hypogene copper mineralization is more closely related to the late argillic type of alteration than to quartz sericitization. Chemical analyses of the altered granodiorite samples showing that the sulfur content of the rock increases considerably in stage 3 support this view.

The evidence suggests that the primary copper mineralization is directly related to an end stage of argillic alteration.

Correlation of Supergene Copper Mineralization with Hydrothermal Alteration

Since chalcocite clearly originated by downward enrichment of a low-grade hypogene protore, there cannot be a direct correlation between chalcocite and the hydrothermal minerals. The distribution of chalcocite in the porphyry must therefore be indirectly controlled by hydrothermal alteration features. Besides an original source of copper (represented in this deposit by the chalcopyrite-bornite protore), two other factors are required for the formation of supergene chalcocite: a sufficiently porous and permeable rock to permit the dilute supergene solutions to reach the zone where secondary sulfides are precipitated, and a mineral on which the descending copper sulfate solutions can precipitate chalcocite (usually chalcopyrite or pyrite). To evaluate these two factors in the development of supergene chalcocite at Santa Rita, the porosity and total iron content of assayed samples of altered granodiorite were determined. The total iron content gives an approximation of the relative amount of pyrite in the rock, except in the least-altered granodiorite specimens where iron is involved principally in the composition of mafic minerals and magnetite. The iron determinations were made at the company assay office at Santa Rita. Porosity of each sample was later obtained in the laboratory by determining the specific gravity of both the bulk and ground specimens.

Porosity represents the percentage of the total volume of the rock that consists of open space, while permeability measures the rate of flow of a fluid through a porous medium. Ohle (1951, p. 671) discusses attempts to calculate permeability from porosity, but the two are not directly related. The porosity determinations were made because of their comparative simplicity and the small amount of equipment required. Although porosity is not an exact measure of permeability, it probably does

TABLE 4.—POROSITY, IRON, AND COPPER CONTENTS OF ALTERED GRANODIORITE PORPHYRY

Sample No.	Specific gravity		% Porosity	% Iron	% Copper
	Bulk	Powder			
A-1	2.36	2.69	12.27	3.4	0.09
A-2	2.35	2.67	11.98	3.8	0.23
A-3				3.5	0.10
A-4	2.43	2.63	7.60	3.3	0.07
A-5	2.36	2.62	9.92	3.3	0.06
A-6	2.46	2.73	9.89	3.2	0.09
A-7	2.38	2.72	12.50	3.9	0.07
A-8				3.4	0.06
B-1	1.94	2.42	19.8	1.9	0.16
B-2				2.5	0.72
B-3				2.3	0.46
C-1	2.33	2.71	14.02	3.5	0.25
C-2	2.41	2.68	10.07	4.9	0.63
C-3	2.34	2.68	12.69	3.9	0.22
C-4	2.33	2.69	13.38	5.8	0.22
C-5	2.55	2.72	6.25	4.5	0.31
C-6	2.46	2.70	8.89	6.7	0.25
C-7	2.42	2.66	9.02	4.8	0.22
C-8	2.42	2.71	10.70	4.4	0.34
C-9	2.33	2.66	12.41	5.6	0.40
C-10	2.35	2.69	12.64	5.4	0.20
D-1	2.10	2.47	14.98	1.0	0.43
D-2	2.23	2.61	14.56	1.0	0.25
D-3	2.27	2.40	5.42	0.6	0.22
D-4	2.21	2.58	14.34	0.8	0.25
D-5	2.14	2.59	17.37	0.5	0.43
E-1	2.38	2.66	10.53	1.7	0.53
E-2	2.17	2.57	15.56	2.0	0.56
E-3	2.20	2.61	15.71	1.4	1.15
E-4	2.14	2.64	18.94	1.2	1.18
E-5	2.19	2.64	17.05	1.1	1.33
E-6	2.12	2.66	20.30	1.1	0.94
E-7	2.22	2.72	18.38	1.4	1.30
E-8				1.2	1.58
E-9	2.17	2.56	15.23	1.1	1.49
E-10	2.28	2.67	14.61	1.2	0.96
E-11	2.23	2.79	20.07	1.7	1.61
E-12				1.4	1.19
F-1	2.41	2.74	12.04	2.1	0.91
F-2	2.15	2.68	19.78	1.8	0.93
F-3	2.29	2.67	14.23	2.2	0.85
F-4	2.36	2.74	13.87	2.2	0.84
F-5	2.35	2.66	11.65	1.9	0.99
F-6	2.27	2.71	16.24	2.6	1.98
F-7	2.28	2.64	13.64	1.9	1.00
F-8	2.24	2.69	16.73	2.5	1.67
F-9	2.28	2.64	13.64	2.7	1.21

reflect variations of permeability with the hydrothermal alteration of the porphyry.

To determine the specific gravity of bulk samples, the specimens were weighed in air, coated with paraffin to prevent hydration of clays and other changes in the rock on immersion in water, reweighed in air to obtain the weight of the paraffin coat, and then weighed submerged in water. The buoyant effect of the wax coat was incorporated into the porosity calculations. At first duplicate determinations were made on each specimen, and individual porosity results varied as much as 3 per cent. The chief cause of variation stemmed from differences in the specific gravity of the bulk fragments from the same specimens. To measure the bulk specimens an improved specific-gravity balance was used. The weight limit of the device is fixed at approximately 35 grams. The difference in specific gravity obtained on two chips of one sample would probably be lessened by using specimens of greater volume, but special equipment would be required. Relative changes in porosity produced by rock alteration in many specimens are of importance and were sought in preference to precise measurements of a few samples. Since the percentage variation did not obscure the general trend of porosity changes, the remaining determinations were carried out as before, but not duplicated. The pyknometer method of measuring specific gravity of pulverized samples, on the other hand, gave very consistent results, with variations of only ±0.02. The total iron content and percentage porosity of each specimen from the various areas sampled (A, B, C, D, E, F) are listed in Table 4.

Correlation of Porosity and Total Iron with Copper Content of Altered Granodiorite

The influence of pyrite and porosity on the formation of secondary chalcocite in the Santa Rita deposits may be observed from the data in Figure 16. A few breaks in the porosity curve indicate that bulk specimens were not available for porosity determinations.

The correlation among copper, iron, and porosity in stage 1 is not evident, perhaps because stage 1 represents only slightly altered granodiorite. The total iron assay is derived from mafic minerals and magnetite to a much greater extent than from pyrite, which

is relatively scarce in the least-altered granodiorite. The porosity of this rock is the lowest of all specimens, averaging about 10 per cent.

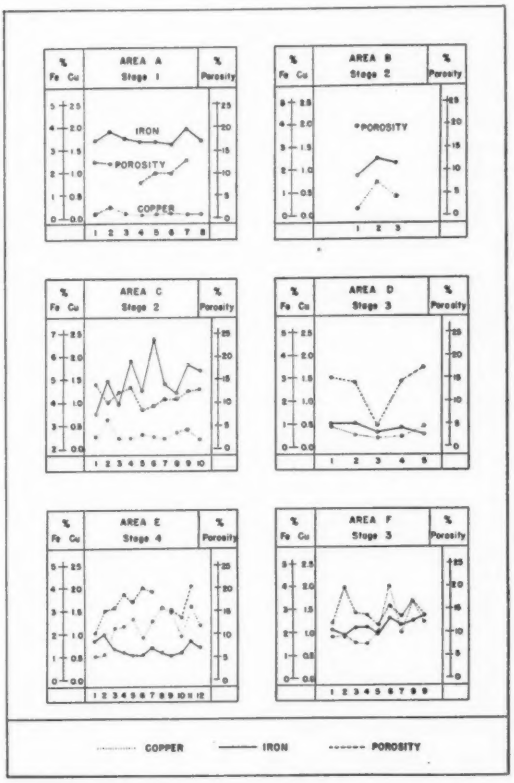

FIGURE 16.—CORRELATION OF POROSITY AND TOTAL IRON WITH COPPER CONTENT OF ALTERED GRANODIORITE PORPHYRY

The copper content is also significantly low. The increase of porosity in one specimen of stage 2 rock suggests that argillization of the montmorillonite-illite type develops much pore spaces. The copper and iron curves of the stage 2 samples correlate nicely and suggest that the copper content occurs as chalcocite coating pyrite. The data from area C yield negative evidence that the distribution of chalcocite is controlled by the presence of pyrite and the porosity of the granodiorite. There is little correlation between copper and porosity in these samples and even less between copper and iron. Adequate explanation for these discrepancies is the nature of the copper occurrence in this set of samples. The polished surfaces of specimens from the siliceous type of stage 2 rock indicate that chalcopyrite with subordinate covellite was the dominant copper mineral. The copper values are thus not the product of supergene enrichment in the form of chalcocite. There is therefore little or no correlation between copper and iron. The lack of correlation is emphasized in individual samples, because pyrite veins are common and raise the iron content. The average assay or total iron for this zone is approximately 5 per cent compared to 3 per cent for the least-altered granodiorite. Though quite siliceous, this rock has a slightly higher porosity than that of stage 1. The brittleness of the rock in area C probably

accounts for the fractures and therefore the higher porosity of the rock. The rock in area D shows the interrelationship among iron, copper and porosity. The poorest copper sample also shows the lowest iron content and the lowest porosity. With an increase in iron and porosity the copper assay is similarly raised. In stage 4 samples the low iron content of the rock (average for the 12 samples 1.4 per cent) reflects the high degree of replacement of pyrite by chalcocite. The strong development of chalcocite, as well as the presence of hypogene chalcopyrite-bornite, is also verified by the relatively high copper values. The average porosity of 16.6 per cent is considerably higher than that of stage 1 samples, and the correlation of copper with porosity is fairly well marked. In the stage 3 rock of area F the influence of porosity and pyrite on supergene chalcocite is again well shown. Porosity is not as high as that in area E, but is still fairly well developed (14.5 per cent), and the similar fluctuation of the copper and porosity curves for a few samples is well shown. The copper and iron curves do not correspond so well.

There seem to be good indications that chalcocite is well developed only where the altered granodiorite contains enough pyrite to control the precipitation of the supergene copper minerals, and where the porosity and permeability of the porphyry are sufficiently high to facilitate downward movement of enriching solutions. The correlation among porosity, pyrite content as reflected by total iron assays, and chalcocite as given by copper assays is not perfect throughout the deposit. Instead of invalidating the general relationship outlined above, the discrepancies suggest that additional factors are involved in the control of supergene enrichment. One might be the influence of structural control on the development of secondary sulfides. This factor has recently been studied at another porphyry copper location (Anderson, 1948, p. 179), and possibly structural control exerted an influence on the formation of the ore body at Santa Rita comparable to that of porosity and pyrite content of the rock.

The increase of porosity with progressive wall-rock alteration is strong indication that the latter results in the former. The development of clays and sericite by hydrothermal solutions, as well as its leaching of certain rock constituents, clearly indicates that increased porosity is to a large extent the work of the pervasive rock-altering solutions. A gain of more than 5 per cent porosity from least- to most-altered granodiorite is apparently to be attributed to these solutions. Undoubtedly increased permeability was more directly affected by the development of fractures and other structural conduits, but this effect cannot be so easily estimated quantitatively.

Pyrite development may not be due entirely to widespread alteration of the granodiorite, although the disseminated pyrite is probably the direct results. In various thin sections marked sulfidation of magnetite in stage 1 rock gives a very plausible and adequate explanation for the dissemination of pyrite in the altered porphyry (Beeson, 1917, p. 384). Some pyrite, however, seems to have originated in another way. The pyrite in the quartz-sericite-pyrite veinlets occurs both as a central, discontinuous seam and as disseminated grains within the halo of sericitic alteration. Probably this pyrite resulted from sulfidation of all rock iron at the sericite front as suggested at Butte by Sales and Meyer (1949, p. 263). Finally, in places pyrite veins up to an inch wide cut altered granodiorite (especially in area C), and these vein fillings clearly represent addition of pyrite to the wall rock from mineralizing solutions. The most significant type of pyrite from the point of view of chalcocite enrichment is undoubtedly the disseminated form produced by sulfidation of magnetite during widespread rock alteration.

In final analysis the stage of hydrothermal alteration of the granodiorite most favorable to chalcocite mineralization appears to be stage 3, for, besides having a high porosity and pyrite content, this stage also contains the source of secondary chalcocite, hypogene chalcopyrite. At the same time, areas of stage 4 rock may also be good ore carriers in view of their well-developed porosity and pyrite contents and more especially if they are, or were at one time, overlain by masses of argillized protore.

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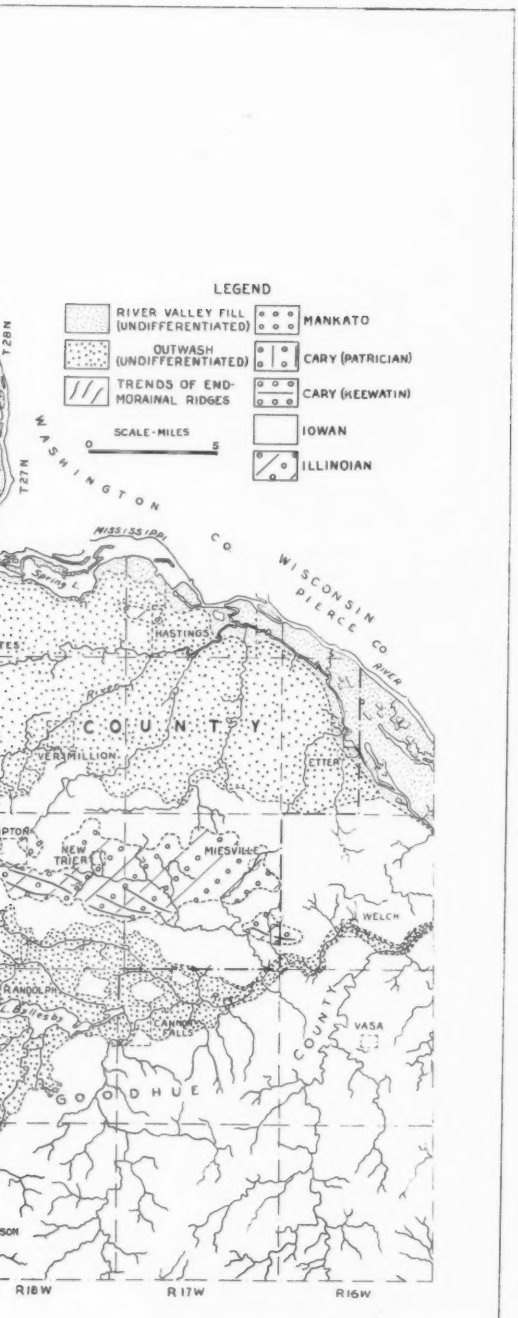
GLACIAL DRIFTS OF THE DAKOTA COUNTY AR



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GLACIAL GEOLOGY OF THE DAKOTA COUNTY AREA, MINNESOTA

By Robert V. Ruhe and Laurence M. Gould

Abstract

Dakota County in southeastern Minnesota provides unique exhibits for the study of Pleistocene drifts derived from north and northwest of the Lake Superior region.

No known Nebraskan till was found, but the Kansan is exposed in road cuts in the central and western parts of the county. The surficial geology consists of tills of Illinoian, Early Wisconsin (Iowan), and Late Wisconsin (Cary of both Keewatin and Patrician origins and Mankato) ages and extensive areas of outwash.

The stratigraphic relationships of the once much-disputed Iowan to the Illinoian and the Illinoian to the Kansan are revealed in road cuts in the west.

Post-Iowan tills are restricted to the north and west. The terms "Keewatin" and "Patrician" are retained to indicate the sources of the Cary drifts.

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Introduction

The area described includes all of Dakota County, Minnesota, and adjacent parts of Goodhue, Rice, and Scott counties (Fig. 1).

FIGURE 1.—LOCATION OF DAKOTA COUNTY AREA (shaded) IN MINNESOTA

To explain the relationships of the glacial drifts, reference must be made to southern Minnesota and adjacent areas in Iowa, South Dakota, and Wisconsin.

Limited studies were made by Gould during 1940 and 1941, and a grant from the Penrose Bequest of the Geological Society of America in 1942 enabled him to spend the summer of 1942 in the field. Gould's work was confined to the Iowan and older drifts.

Under the auspices of a National Research Fellowship in the Natural Sciences, Ruhe did field work during parts of July–December 1950 and April–May 1951. Field expenses were subsidized by the Industrial Science Research Institute (Project 273), Iowa State College. This aid is gratefully acknowledged. The work of Ruhe pertains mainly to the Keewatin Cary and younger drifts.

Previous Studies

Earlier studies of counties of southeastern Minnesota, mainly the work of Winchell, Upham, and Harrington, were published in final reports of the Minnesota Geological and Natural History Survey (vol. 1, 1884; vol 2, 1888). Sardeson (1916) described the northernmost part of the area. Leverett and Sardeson (1919) discussed the whole of southeastern Minnesota. Leverett (1932), with contributions by Sardeson, published a detailed account of the Quaternary geology of Minnesota. Cooper's work (1938), although pertaining mainly to the Pleistocene geology north of Minneapolis and St. Paul, influences drift correlations south of the Twin Cities. Schwartz (1936) described the geology of the metropolitan area of Minneapolis and St. Paul and included a part of the present area. Gould (1941) described the relationships of the Iowan and Illinoian drifts in Dakota County, Minnesota. Thiel (1944) discussed the surficial deposits of southeastern Minnesota in relationship to his ground-water studies. In Iowa, the Pleistocene correlations are described by Kay and Apfel (1929) and Kay and Graham (1943). Ruhe (1950a; 1950b) reclassified and correlated the glacial drifts of northwestern Iowa and southwestern Minnesota. Interpretations of the drifts of that region bear directly on the drift correlations of north-central and northeastern Iowa, southeastern Minnesota, and the Dakota County area.

Pre-Wisconsin Drifts

Nebraskan Drift

Leverett (1932, p. 14–20) summarized the scant information available on the Nebraskan drift in the Dakota County area and southeastern Minnesota. The writers have not been able to identify a Nebraskan drift in the area.

Kansan Drift

Areal distribution.—Glacial drift that may be Kansan is exposed at several places in Dakota County. Sections occur several miles northwest of Farmington, near Hampton, and near Miesville. The drift is yellowish brown, contains rock types from sources to the west and northwest of Dakota County, and is overlain by a red drift of Lake Superior type lithology. The red drift is the "old red" Leverett tentatively classified as Illinoian.

Although exposures of the lower yellowish-brown drift are few and widely separated

geographically, it is reasonable to postulate that the whole Dakota County area was subjected to the glacier ice that deposited this till. Classification and correlation of the drift as Kansan is problematical. (*See* section on Stratigraphy.)

Topography.—The Kansan drift is buried under younger deposits and does not influence the topography. Schwartz (1936, p. 65–80) has shown that the bedrock surface in this area may have up to 500 feet of relief. Configuration of the bedrock surface appears to control the topographic expressions of the pre-Cary drifts. The Kansan drift occupies both upland and valley-slope positions. For example, in the NW¼ sec. 22, T.114 N., R.20 W. (Section 1), and in the ECSE¼ sec. 17, T.113 N., R.18 N. (Section 1), the yellowish-brown till overlain by red drift was observed on upland ridge crests. In the latter locality, the Kansan rests on St. Peter sandstone. However, in the EC sec. 14, T.114 N., R.20 W. (Section 1) the same till, overlain by red, outcrops on a valley slope 80 to 100 feet below the upland.

Lithology.—The most significant aspect of the Kansan till in Iowa is the intensely weathered surface zone (Kay and Apfel, 1929, p. 235–253). Two weathered phases of the drift, the gumbotil and the feretto, long have been recognized as important horizons that represent weathering during interglacial times and that separate superimposed drifts. Simonson (1941, *passim*) and Scholtes, Ruhe, and Riecken (1951, *passim*) have pointed out that the gumbotil in Kansan till possesses the morphological characteristics of the B horizon of a Planosol developed under grass or forest vegetation or both. Scholtes *et al.* also showed that the feretto zone is the B horizon of a soil in Kansan till that has the morphological characteristics of a Gray-Brown Podsol probably formed under forest vegetation. Both papers indicate that the upper part of the Kansan drift has been thoroughly weathered and decomposed.

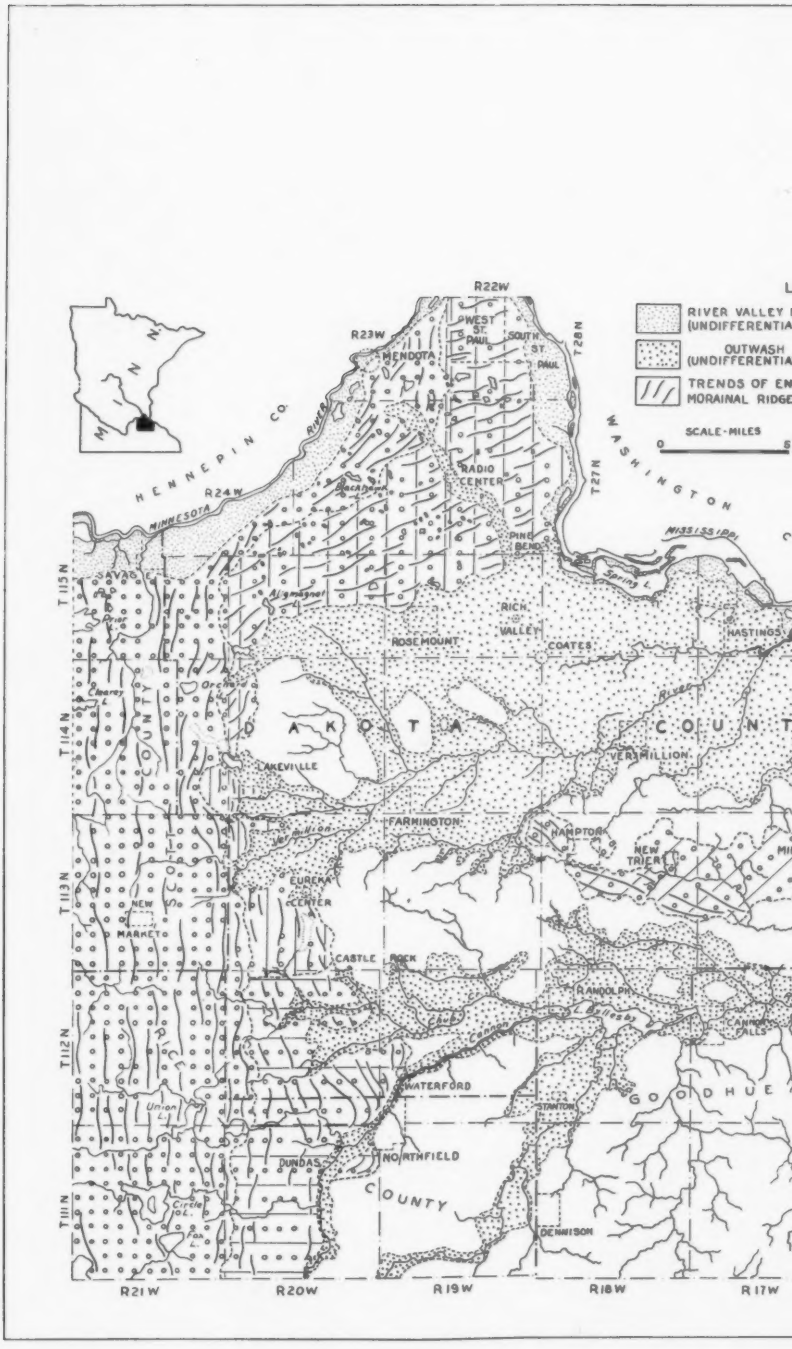
In Dakota County, the gumbotil or feretto has not been found in the upper part of the drift thought to be Kansan. In three sections (Section 1), the yellowish-brown till under the old red drift is leached to depths as much as 8 feet. Pebbles in the leached zone (K-1, K-2; Table 1) show only slight decomposition. Pink granite pebbles are the dominant rock type. Basalt pebbles are also abundant. These constituents are abundant in the pebble suite of the calcareous drift phase of the Kansan (K-3; Table 1). The only apparent difference between the leached and calcareous phases of this drift in Dakota County is the presence of limestone pebbles in the latter. Composition of the leached phase of the Kansan of this area differs markedly from the pebble content of the Kansan gumbotil of Iowa where siliceous pebbles, quartz, and chert are the dominant rock types (Kay and Apfel, 1929, p. 238). Content of the leached phase is similar to the leached Kansan drift phase of the gumbotil area (Kay and Apfel, 1929, p. 245).

At one locality, the EC sec. 14, T.114 N., R.20 W., Dakota County (K-3), calcareous Kansan till is exposed. Limestone pebbles constitute 42 per cent of the pebble assemblage. In the road cut, calcareous Kansan is overlain by calcareous old red drift.

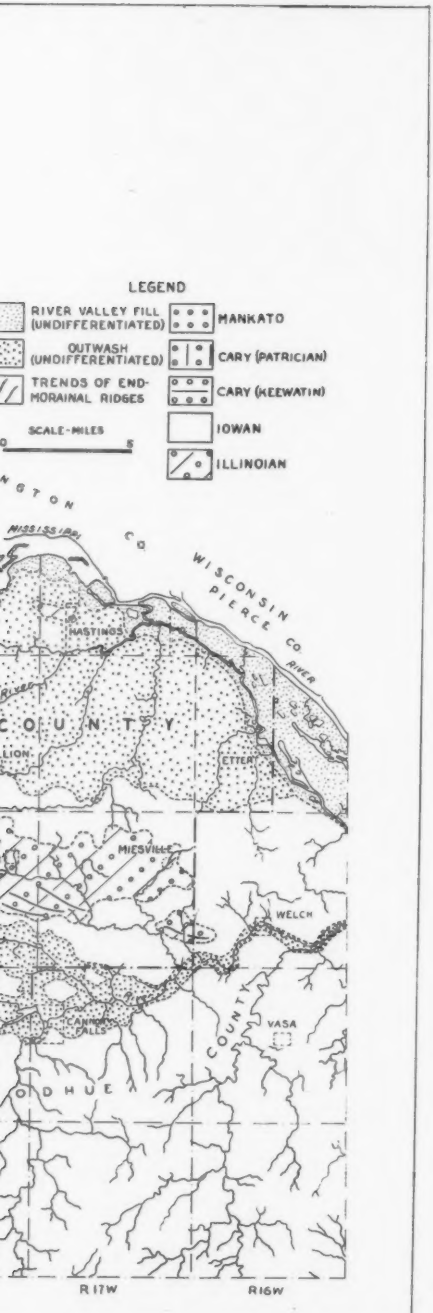
The pebble analyses show significant amounts of red sandstone and iron-formation pebbles in the Kansan till (K-2, K-3; Table 1). Red sandstone and iron-formation pebbles are common in the Illinoian and Patrician Cary drifts in Dakota County. Source of the rock in these two drifts is the Lake Superior region. Lake Superior type rock in the Kansan drift requires that depositing ice advanced: (1) from the north, or (2) from the west and northwest but over materials, glacial or otherwise, derived from the Lake Superior region, or (3) over unknown outcrops of the Lake Superior type rock to the northwest. Number (1) can be discarded because the bulk of the Kansan drift is of northwestern provenance, and the texture is similar to all other drifts of northwestern origin and differs markedly from that of the drifts derived in the north and northeast (Fig. 2).

Although a "red" drift has not been found under the Kansan drift, its occurrence in this area is a possibility. For example, the Keewatin Iowan drift that overlaps the Patrician Illinoian red drift contains minor amounts of Lake Superior type rock.

The Kansan till is a silty clay (Fig. 2). Plots of analyses of samples fall within the drifts of western and northwestern provenance. All these



GLACIAL DRIFTS OF THE DAKOTA COUNTY AREA, MINNE



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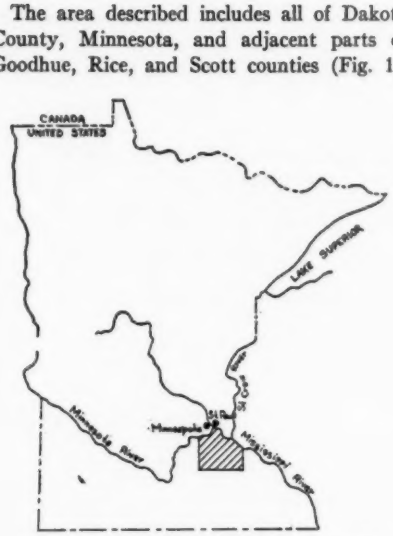

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Pre-Wisconsin Drifts

Nebraskan Drift

Leverett (1932, p. 14–20) summarized the scant information available on the Nebraskan drift in the Dakota County area and southeastern Minnesota. The writers have not been able to identify a Nebraskan drift in the area.

Kansan Drift

Areal distribution.—Glacial drift that may be Kansan is exposed at several places in Dakota County. Sections occur several miles northwest of Farmington, near Hampton, and near Miesville. The drift is yellowish brown, contains rock types from sources to the west and northwest of Dakota County, and is overlain by a red drift of Lake Superior type lithology. The red drift is the "old red" Leverett tentatively classified as Illinoian.

Although exposures of the lower yellowish-brown drift are few and widely separated

geographically, it is reasonable to postulate that the whole Dakota County area was subjected to the glacier ice that deposited this till. Classification and correlation of the drift as Kansan is problematical. (*See* section on Stratigraphy.)

Topography.—The Kansan drift is buried under younger deposits and does not influence the topography. Schwartz (1936, p. 65–80) has shown that the bedrock surface in this area may have up to 500 feet of relief. Configuration of the bedrock surface appears to control the topographic expressions of the pre-Cary drifts. The Kansan drift occupies both upland and valley-slope positions. For example, in the NW¼ sec. 22, T.114 N., R.20 W. (Section 1), and in the ECSE¼ sec. 17, T.113 N., R.18 N. (Section 1), the yellowish-brown till overlain by red drift was observed on upland ridge crests. In the latter locality, the Kansan rests on St. Peter sandstone. However, in the EC sec. 14, T.114 N., R.20 W. (Section 1) the same till, overlain by red, outcrops on a valley slope 80 to 100 feet below the upland.

Lithology.—The most significant aspect of the Kansan till in Iowa is the intensely weathered surface zone (Kay and Apfel, 1929, p. 235–253). Two weathered phases of the drift, the gumbotil and the feretto, long have been recognized as important horizons that represent weathering during interglacial times and that separate superimposed drifts. Simonson (1941, *passim*) and Scholtes, Ruhe, and Riecken (1951, *passim*) have pointed out that the gumbotil in Kansan till possesses the morphological characteristics of the B horizon of a Planosol developed under grass or forest vegetation or both. Scholtes *et al.* also showed that the feretto zone is the B horizon of a soil in Kansan till that has the morphological characteristics of a Gray-Brown Podsol probably formed under forest vegetation. Both papers indicate that the upper part of the Kansan drift has been thoroughly weathered and decomposed.

In Dakota County, the gumbotil or feretto has not been found in the upper part of the drift thought to be Kansan. In three sections (Section 1), the yellowish-brown till under the old red drift is leached to depths as much as 8 feet. Pebbles in the leached zone (K-1, K-2; Table 1) show only slight decomposition. Pink granite pebbles are the dominant rock type. Basalt pebbles are also abundant. These constituents are abundant in the pebble suite of the calcareous drift phase of the Kansan (K-3; Table 1). The only apparent difference between the leached and calcareous phases of this drift in Dakota County is the presence of limestone pebbles in the latter. Composition of the leached phase of the Kansan of this area differs markedly from the pebble content of the Kansan gumbotil of Iowa where siliceous pebbles, quartz, and chert are the dominant rock types (Kay and Apfel, 1929, p. 238). Content of the leached phase is similar to the leached Kansan drift phase of the gumbotil area (Kay and Apfel, 1929, p. 245).

At one locality, the EC sec. 14, T.114 N., R.20 W., Dakota County (K-3), calcareous Kansan till is exposed. Limestone pebbles constitute 42 per cent of the pebble assemblage. In the road cut, calcareous Kansan is overlain by calcareous old red drift.

The pebble analyses show significant amounts of red sandstone and iron-formation pebbles in the Kansan till (K-2, K-3; Table 1). Red sandstone and iron-formation pebbles are common in the Illinoian and Patrician Cary drifts in Dakota County. Source of the rock in these two drifts is the Lake Superior region. Lake Superior type rock in the Kansan drift requires that depositing ice advanced: (1) from the north, or (2) from the west and northwest but over materials, glacial or otherwise, derived from the Lake Superior region, or (3) over unknown outcrops of the Lake Superior type rock to the northwest. Number (1) can be discarded because the bulk of the Kansan drift is of northwestern provenance, and the texture is similar to all other drifts of northwestern origin and differs markedly from that of the drifts derived in the north and northeast (Fig. 2).

Although a "red" drift has not been found under the Kansan drift, its occurrence in this area is a possibility. For example, the Keewatin Iowan drift that overlaps the Patrician Illinoian red drift contains minor amounts of Lake Superior type rock.

The Kansan till is a silty clay (Fig. 2). Plots of analyses of samples fall within the drifts of western and northwestern provenance. All these

TABLE 1.—PEBBLE ANALYSES OF GLACIAL DRIFTS

(In per cent by count)

Rock type	K-1	K-2	K-3	Il-1	Il-2	Il-3	Il-4	Il-5	I-1	I-2	I-3	I-4	Ck-1	Ck-2	Ck-3	Ck-4	Cp-1	Cp-2	Cp-3	Cp-4	Cp-5	Cp-6	Cp-7	Cp-8	M-1	M-2	M-3
Granite	43	15	40	14	21	15	25	10	26	12	19	21	18	26	17	17	24	22	20	21	22	15	21	15	32	31	24
Quartz	34	6	19	34	11	11	37	31	11	3	4	8	..	3	2	4	1	3	..	Tr	5	2	2	4	..	..	1
Chert	8	3	4	..	6	4	3	12	2	Tr	Tr	5	2	1	2	1	4	6	..	..	2	2	2	7	..	1	2
Basalt	7	15	8	10	19	27	15	17	15	20	18	23	23	7	21	10	30	19	35	39	17	44	47	34	23	12	19
Quartzite	4	1	3	3	4	4	1	1	3	1	3	1	5	..	3	1	..	..	..	16	3	10	7	14	8	4	1
Sandstone	2	8	18	21	15	23	3	10	14	7	11	12	3	2	4	6	18	24	27	9	3	5	9	5	4	5	5
Schist	1	2	2	8	4	Tr	1	2	Tr	..	..	3	..	..	1	..	6	2	3	4	..	..	1	1	..	..	..
Felsite	1	1	..	1	1	..	..	..	3	..	Tr	..	..	..	..	1	..	3	1	..	Tr	..	Tr	1	..	..	..
Limestone	..	42	..	..	..	11	..	..	18	55	41	25	45	59	36	60	..	..	..	..	41	12	4	..	27	43	32
Graywacke	..	3	1	..	6	2	4	7	3	..	..	..	..	..	2	..	2	7	..	Tr	..	..	..	..	..	..	1
Iron fm.	..	1	2	..	1	..	1	2	Tr	..	..	..	..	..	..	..	2	..	..	1	..	..	Tr	1	..	..	..
Aplite	..	1	..	..	..	..	..	..	..	..	..	..	..	..	..	..	2	3	..	..	..	3	1	..	..	..	..
Trachyte	..	1	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	Tr	..	..	..	..	..	..
Gabbro	..	1	..	1	1	Tr	1	2	1	Tr	..	1	..	1	..	..	1	5	8	4	3	7	2	8	1	1	..
Greenstone	..	1	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
Siltstone	..	..	2	..	7	..	5	6	..	..	..	..	..	..	..	1	..	..	..	..	..	..	..	..	1	..	..
Syenite	..	..	..	..	2	Tr	1	..	1	..	..	1	..	..	..	..	7	..	3	2	Tr	..	Tr	8	..	..	..
Porphyry	..	..	..	2	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	Tr	..	..	..	..	..	..	..
Diorite	..	..	..	4	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	1	..	..	..	..	..
Shale	..	..	..	..	..	..	1	..	..	Tr	1	..	4	1	11	..	..	1	3	1	..	Tr	Tr	1	5	3	15
Conglomerate	..	..	..	..	..	1	1	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
Rhyolite	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	2	..	..	..	..	Tr	Tr	..	..	..	..
Gneiss	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	1	2	..	..	..	1	..	..	..	..	..
Arkose	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	1	..	Tr	..	..	..	..	..	..	..

K—Kansan, Il—Illinoian, I—Iowan, Ck—Keewatin Cary, Cp—Patrician Cary, M—Mankato, Tr—less than 1 per cent

Il-3: Illinoian till, EC sec. 14, T.114N., R.20W., Dakota County
Il-4: Illinoian gravel, NC sec. 16, T.111N., R14W., Goodhue County
Il-5: Illinoian gravel, C sec. 16, T.111N., R.14W., Goodhue County
I-3: Iowan till, NW1/4 sec. 27, T.114N., R.20W., Dakota County
I-4: Iowan till, NE1/4 sec. 16, T.113N., R.18W., Dakota County
CK-1: Keewatin Cary till, SW cor. sec. 33, T.113N., R.20W., Dakota County
CK-3: Keewatin Cary till, NE cor. sec. 25, T.112N., R.20W., Dakota County
CK-4: Keewatin Cary till, WC SW1/4 sec. 29, T.113N., R.20W., Dakota County

See text for other samples.

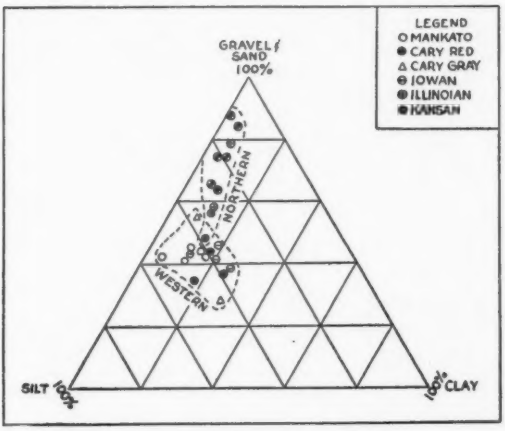


FIGURE 2.—TEXTURES OF TILLS

drifts—Kansan, Iowan, Keewatin Cary, and Mankato—were deposited by ice associated with the Keewatin "center". All have similar pebble assemblages and similar textures.

Stratigraphy.—Stratigraphic relationships of the yellowish-brown drift (Kansan?) to its overlying old red drift are shown in the exposures of section 1.

SECTION 1

K-1: NW¼ sec. 22 T. 114 N., R. 20 W., Dakota County (road cut), (Fig. 1; Pl. 3)

	Thickness Feet	Inches
4. Iowan and post-Iowan loess:	9	0
Yellowish brown, well sorted, leached	5	6
Yellowish brown, well sorted, calcareous	3	6
3. Iowan till: (I-1, Table 1)	4	6
Yellowish brown, friable, silty, leached	3	3
Yellowish brown; friable, silty, calcareous	1	3
2. Illinoian(?) gravel: red, leached, poorly sorted (Il-1, Table 1)	5	6
1. Kansan(?) till: yellow-gray, compact, plastic, leached (K-1, Table 1)	8	0

K-2: ECSE¼ sec. 17, T. 113 N., R. 18 W., Dakota County (railroad cut 1 mile south of Hampton)

	Thickness Feet
5. Iowan and post-Iowan loess: yellowish brown, leached	1–2
4. Iowan till: yellow, sandy, friable, leached, west cut (I-2, Table 1)	2–3
3. Illinoian(?) till: red, sandy, gravelly, locally, compact and plastic locally, noncalcareous (I1-2, Table 1)	7–8
2. Kansan(?) till: dark yellowish brown, compact, plastic, leached, exposed 75 feet north of reddish yellow St. Peter sandstone band in center of east cut (K-2, Table 1)	6–7
1. St. Peter sandstone: varicolored, friable	15–20

K-3: sec. 14, T. 114 N., R. 20 W., Dakota County (road cut)

	Thickness Feet	Inches
3. Iowan till: gray, calcareous, both primary and secondary carbonate (I-3, Table 1)	2	2
2. Illinoian(?) till: red, sandy, friable, calcareous (Il-3, Table 1)	3	8
1. Kansan(?) till: yellowish brown, compact, calcareous (K-3, Table 1)	4	0

K-4: SW¼ sec. 11, T. 111 N., R. 17 W., Dakota County (road cut)

	Thickness Feet
2. Illinoian(?) till: red, leached, gravelly	1–2
1. Kansan(?) till: yellow-gray, compact, leached	4–5

The contact between the Kansan and the overlying Illinoian in the sections shows no significant break that may represent an interglacial age. Where the Kansan is leached, the overlying Illinoian is leached or may have been originally noncalcareous. Where the Kansan is calcareous, the overlying Illinoian is calcareous. The absence of a soil profile may be explained on the basis of truncation of the solum of the profile in the Kansan drift prior to deposition of the Illinoian drift. Although bedrock relief is great in this area (up to 500 feet locally)

and slopes may have remained steep following Kansan glaciation with resultant removal of the solum, two of the sections are on uplands where slopes are not steep. Could these drifts be closely related in time, one from the Keewatin center and the other from the Patrician? It will be shown that such was the case in the late Wisconsin, but further study of this problem is necessary.

Illinoian Drift

Areal distribution.—Distribution of a drift that may be Illinoian is confined mainly to southeastern Dakota County (Pl. 1; *cf.* Leverett, 1932, Pl. 2), Here the old red is the surface drift.

Limits of the Illinoian drift are difficult to map. Outside the main area near Hampton, there are: (1) a small patch in T.115 N., R.17 W., and (2) two other isolated exposures approximately 10 miles northwest of the main body of the drift (Section 1). Winchell (1888, p. 48) described boulders including amygdaloidal rock, flesh-red prophyry, dolerite, and native copper in the vicinity of Zumbrota and Wanamingo, Goodhue County. These towns are approximately 16 to 20 miles south and southeast of the old red drift area in southeastern Dakota County. Winchell (1884, p. 648) had located previously a red till in the NW¼ sec. 9, Cannon City Township, Rice County. Pebbles and boulders of red quartz prophyry were thought to be derived from the copper-bearing series of the northeast part of the State. He also reported native copper from these localities. Red till overlying St. Peter sandstone was observed by Winchell in NE¼ sec. 5 of this same township. Leverett and Sardeson (1919, p. 110–111) reported red drift along the tracks of the Rock Island Railroad in NC, sec. 5, Cannon City Township. They interpreted this red drift, however, as masses that had been transported south by the Keewatin ice that deposited the young gray drift (Leverett's Late Wisconsin). The Cannon City Township sections are approximately 22 miles south of the Illinoian drift area in southeastern Dakota County.

Three exposures in Goodhue County suggest the occurrence of Illinoian drift.

SECTION 2

Il-1: C sec. 33, T. 112 N., R. 17 W., Goodhue County (road cut)

	Thickness Feet
2. Iowan and post-Iowan loess: yellowish brown, well sorted; upper 5½ feet leached	12
1. Illinoian(?) outwash: reddish brown, leached sand and gravel; contains red sandstone pebbles	1½

Il-2: NC sec. 16, T.111 N., R.14 W., Goodhue County (road cut)

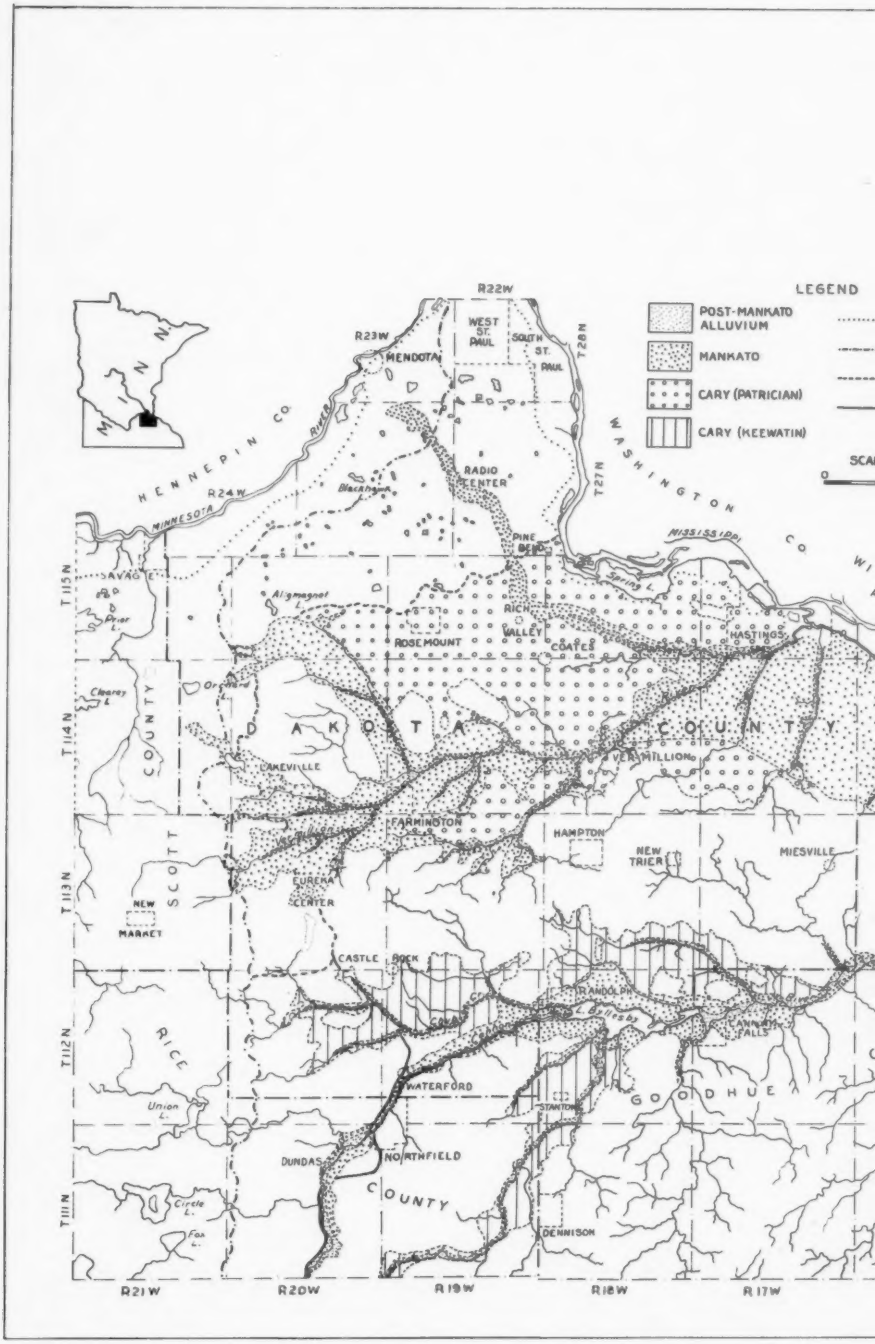
	Thickness Feet
2. Iowan and post-Iowan loess: yellowish brown, calcareous	17
1. Illinoian(?) outwash: reddish brown, leached sand and gravel; red sandstone and iron formation pebbles (Il-4, Table 1)	2

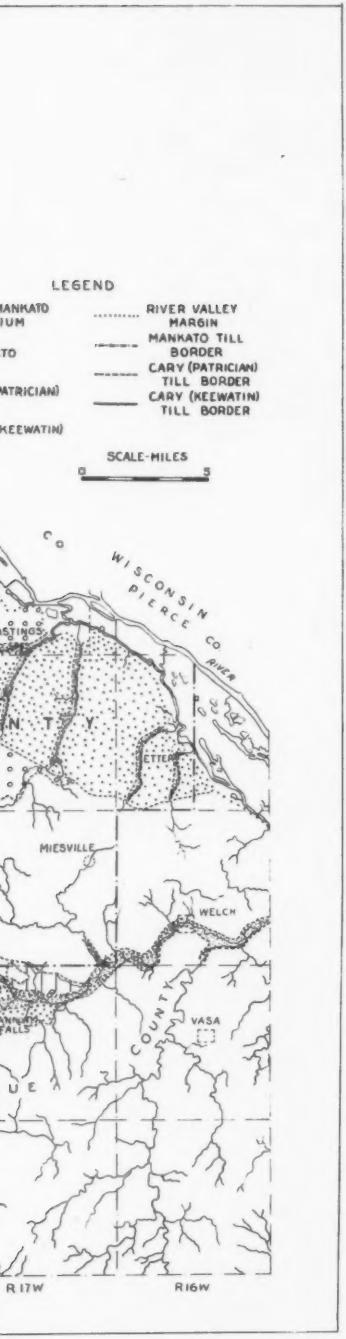
Half a mile south of Il-2, a similar gravel (Il-5, Table 1) is exposed under 6 feet of loess. These gravels are on the uplands and are not related to present drainage courses. They are probably remnants of an Illinoian outwash apron associated with the Hampton morainic system, which if projected southeastward trends through this area. If these gravels were reworked during Wisconsin time, they should be related to drainage ways on the Iowan surface, but they are not.

Leverett (1932, Pl. 2) mapped a line that roughly corresponds geographically to the trend of the Hampton moraine. It is not clear whether this line signifies the location of the moraine or the limit of the Illinoian drift. Winchell's reports and the sections observed by the writers indicate that the limit of the Illinoian red drift is not marked by the Hampton moraine, but that areas farther south and west were covered by the Illinoian ice.

Topography.—The Illinoian is the oldest drift in southeastern Minnesota that has an end moraine (Pl. 1). Slopes within this topographic belt range up to 40 per cent, and relief up to 100–150 feet. Trends of ridges are identifiable on aerial photographs and are readily traced in the field.

In places the morainic expression of these ridges appears to reflect high bedrock (Fig. 1; Pl. 4). Yet, some road cuts through ridges show Illinoian drift several tens of feet thick. In the SE¼ sec. 25, T.113 N., R.18 W., Dakota County, a road cut exposes 25½ feet of non-


LEGEND
POST-MANKATO ALLUVIUM
MANKATO
CARY (PATRICIAN)
CARY (KEEWATIN)
MINN.
R22W
R23W
R24W
T28N
T27N
T115N
T114N
T113N
T112N
T111N
R21W
R20W
R19W
R18W
R17W
WEST ST. PAUL
SOUTH ST. PAUL
MENDOTA
HENNEPIN CO.
WASHINGTON CO.
RADIO CENTER
PINE BEND
MINNESOTA RIVER
MISSISSIPPI
SAVAGE
Prior L.
Cleary L.
Alimagnet L.
Orchard L.
Spring L.
RICH VALLEY
ROSEMOUNT
COATES
HASTINGS
DAKOTA COUNTY
LAKEVILLE
VERMILLION
FARMINGTON
HAMPTON
NEW TRIER
MIESVILLE
SCOTT COUNTY
NEW MARKET
EUREKA CENTER
CASTLE ROCK
RANDOLPH
CANNON FALLS
RICE COUNTY
WATERFORD
Union L.
GOODHUE COUNTY
STANTON
DUNDAS
NORTHFIELD
DENNISON
Circle L.
Fox L.


LEGEND
RIVER VALLEY MARGIN
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CARY (KEEWATIN) TILL BORDER
SCALE-MILES
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5
WISCONSIN
PIERCE CO
RIVER
ETTER
MIESVILLE
WELCH
VASA
COUNTY
R 17W
R 16W

A, MINNESOTA

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calcareous red gravel (section Il-3). Although bedrock is high, the linear ridges of the Hampton moraine are a result of glacial construction (Fig. 2; Pl. 4).

The end-morainal surface is well drained. Closed depressions are lacking. This end moraine has maintained its constructional features, somewhat modified, since Illinoian time. The slopes and relief in this area imply that the moraine should be subjected to severe erosion by subaerial processes, but that is not the case. The lack of severe erosion probably is explained by the texture of the drift. Numerous kame ridges and sandy till ridges occur in this morainic belt. Water seeps into the land forms rather than running off the steep slopes.

Distribution of the Illinoian ground moraine is shown on Plate 1. Slopes in this area range up to 8 per cent, and relief averages 40–60 feet. A typical expression of the Illinoian ground moraine occurs along a traverse from New Trier to Miesville. Many road cuts in this area show sections of noncalcareous red Illinoian till up to 8 feet thick.

The ground moraine is well drained. This surface is strikingly similar to the Iowan drift surface. Mapping of limits of the Illinoian till is based on outcropping of the red till in contrast to the yellow Iowan till that surrounds the red-till inlier.

Lithology.—The red to reddish-brown till differs distinctly from the yellow to gray tills of provenance to the northwest. Color presumably is the result of the pulverization and dissemination of ferruginous material throughout the till matrix. Below the zone of the surficial soil profile, where a change in color may occur, red predominates to the base of all sections observed. In several cuts the Illinoian red drift is 25 feet or more thick.

Texture of the till varies but, in general, differs from the "northwestern" tills. The old red is sandy to gravelly, but locally may be of a sandy-silt texture (Fig. 2). A twofold textural division is apparent on the diagram—a northern group and a western group. The northern group embodies tills deposited by ice from the Patrician "center", whereas the western group is the result of Keewatin ice deposition. The Illinoian samples fall within the northern group.

Pebble analyses of Illinoian drift samples showed the lithology to be a Lake Superior type (Table 1).

Stratified drift is abundant in the Hampton moraine. Kames are numerous. Section Il-3 showed 25 or more feet of gravelly material. Remnants of Illinoian outwash (Il-1, Il-2) are believed to mantle the uplands in Goodhue County.

Sardeson (1933, p. 264) reported an outwash fan sloping southward from the Hampton moraine to the juncture of Chub Creek and Cannon River and spreading westward up Chub Creek in Dakota County and Prairie Creek in Dakota and Goodhue counties. Illinoian outwash may be present, but if so it is buried beneath younger deposits. The distal slope of the Hampton moraine is overlapped by Iowan drift. The outwash exposed along Chub and Prairie creeks is part of the valley trains related to a much younger drift, the Cary of the Des Moines lobe.

Loess relationships of Dakota County are not clear. A yellowish-brown, leached to calcareous silt mantles all drifts from Illinoian to late Wisconsin (Patrician Cary). The yellowish loess covering the Illinoian ground moraine is probably Iowan and post-Iowan. Above the Illinoian till the loess is thin and has been leached of carbonate so that the normal stratigraphic section shows leached yellowish-brown loess over leached red Illinoian till. In Goodhue County (Il-1, Il-2) the loess is calcareous above noncalcareous Illinoian gravel. In sections (K-3) where calcareous Iowan till overlies leach Illinoian till, the yellowish-brown loess is above the Iowan.

No loess was found that is directly related to the Illinoian. In Iowa and Illnois, the Loveland loess (Leighton and Willman, 1950, p. 601–602) is believed correlative of the Illinoian. Loveland loess was not identified in Dakota County.

Stratigraphy—Several sections (K-1, K-2, K-3, Il-1, Il-2) illustrate the stratigraphic relationships of the Illinoian drift to the Kansan and Iowan drifts, and post-Iowan loess.

No sections were observed in which calcareous Illinoian drift rests on weathered Kansan drift. Calcareous Iowan till and Iowan

loess were observed over leached or noncalcareous Illinoian drift.

Problem of the Classification and Correlation of the Kansan and Illinoian Drifts

Sardeson (1933, p. 263) credits R. T. Chamberlin with recognition in 1906 of the red drift of southeastern Dakota County, Minnesota, as a distinct glacial drift, and also credits Leverett with recognition of the age of this drift as Illinoian. Actual classification of the old red drift as Illinoian appeared in the literature in 1919 (Leverett and Sardeson, p. 48–49). The old red drift was classified as such because: (1) it lies on the eroded surface of the Kansan (Leverett and Sardeson, 1919, p. 104–106), and (2) it is more weathered than the Wisconsin drift and corresponds in degree of weathering to the Illinoian drift of Illinois. The authors gave no supporting evidence for these statements. Present evidence is not sufficient to affirm absolutely the classification of the old red drift as Illinoian and the underlying yellowish brown till as Kansan.

No sections were observed to show the Patrician drift (Illinoian of Leverett) above a maturely weathered soil profile on the Keewatin till (Kansan of Leverett) such as is found in southeastern Iowa (Kay and Apfel, 1929, p. 226–228; Kay and Graham, 1943, p. 24–31). In two localities (K-1, K-2) leached or noncalcareous Patrician red drift lies on oxidized and leached Keewatin till. A solum, if it had developed in the Keewatin till, may have been stripped prior to deposition of the Patrician drift. On the other hand, the thin Patrician drift may have been deposited on the Keewatin prior to weathering of the latter, and the whole sequence leached of carbonate contemporaneously. Section K-3 showed calcareous Patrician red till on calcareous Keewatin till. The solum and oxidized and leached Keewatin till may have been stripped prior to deposition of the calcareous red till, or the section may represent a depositional sequence without a marked break between the two phases.

Leverett (1932, p. 22) stated that the Illinoian drift blankets the eroded surface of the underlying drift (Kansan), covering the valley bottoms and slopes as well as the uplands. In the area of the Patrician red drift in Dakota County, relief on the bedrock surface ranges up to several hundreds of feet. The eroded surface of the bedrock is probably preglacial (Schwartz, 1936, p. 65–80). Keewatin drift under Patrician has been identified as far east as Miesville (K-4) so it is known that this area of rugged bedrock relief was subjected to a pre-Patrician Keewatin glaciation. There is no evidence that the bedrock valleys were filled by Keewatin drift to or near the levels of uplands, and were then eroded to a surface similar to the present topography. Leverett's observation that the Illinoian covers valley bottoms, slopes, and uplands was based upon relation of observed positions of the drift to the modern landscape. The present writers believe that the bottom land-slope-upland positions of the Patrician drift is more a relection of the bedrock surface than superposition on an erosional surface in Keewatin drift. The Keewatin drift beneath the Patrician and the Iowan drift above the Patrician have been observed in all these topographic positions. Bedrock surface appears to be the dominant control of the topographic positions of the drifts in this area. Thus, although Leverett believed that the Illinoian drift lies on the eroded surface of the Kansan, other interpretations of the same evidence contradict his conclusion.

Leverett's (1919, p. 48–49) observation that the weathering of the Illinoian drift of Dakota County corresponds in degree to the weathering of the Illinoian drift of Illinois is questionable. Leighton and MacClintock (1930, *passim*) described maturely weathered profiles (siltil, mesotil, and gumbotil) in the Illinoian till in Illinois. Comparable weathered profiles have not been observed in Dakota County in the Patrician red drift. Scholtes *et al.* (1951, p. 299) call attention to the difficulties in regional correlation of drifts on the basis of the magnitude of the soil profile (*i.e.*, degree of weathering) developed on the drift surfaces. Time is the factor necessitating evaluation in a correlation of the Patrician red drift of Dakota County and the Illinoian drift of Illinois. Granting topographic control of sample sites, specific controls of parent material, climate, and biotic factors are necessary to compare the ages of the drifts of the two regions. Variability of these factors prevents any significant assessment of weathering values that would lead to correlation.

The contact between the Patrician red drift and the underlying Keewatin drift does not permit age differentiation of these two drifts. The stratigraphic break between the Patrician red drift and the overlying Iowan till, the surface drift that can be traced directly southward through Minnesota into the type Iowan area in Iowa, is more distinct. Although sections K-2 and K-3 show leached Iowan over noncalcareous Patrician, respectively, section K-1 shows calcareous Iowan till over leached Patrician drift. Then, too, calcareous Iowan loess that mantles calcareous Iowan till has been observed over leached Patrician gravels (Il-1, Il-2). The weathered zone between Iowan till and loess and the underlying Patrician drift probably differentiates the Iowan and Patrician on an age basis.

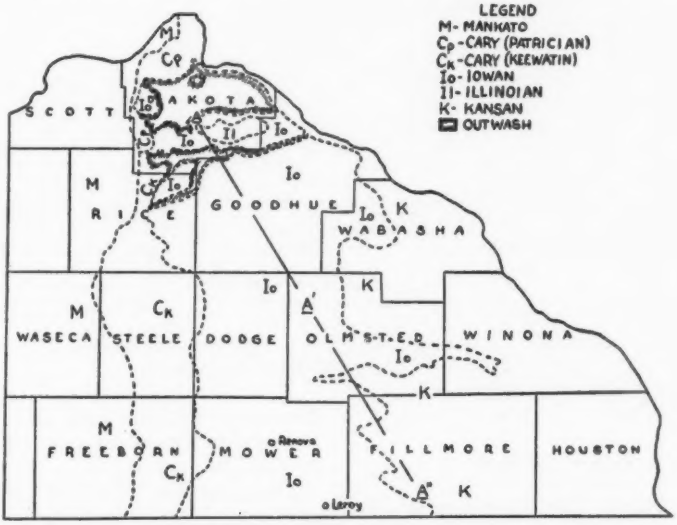

FIGURE 3.—DRIFT DISTRIBUTIONS IN SOUTHEASTERN MINNESOTA

Stratigraphic relationships of the Iowan, Patrician red, and Keewatin yellowish-brown tills may be summarized: (1) the Iowan is separated from the underlying Patrician by a weathered zone in the Patrician drift that may represent an age differentiation, but (2) the Patrician red drift is not separated from its underlying Keewatin drift by a soil or weathering profile. Solution of the relationships of (2) may be found in southeastern Minnesota south of Dakota County.

In Mower County (Fig. 3) the Iowan drift has been observed over Kansan gumbotil near Leroy and Renova. East of the Iowan drift border in Fillmore, Olmsted, and Wabasha counties, Iowan and post-Iowan loess mantles a maturely weathered profile in Kansan till—*i.e.*, a drift that can be traced southward into the recognized Kansan of northeastern Iowa. The Kansan till contains rocks derived from the igneous and sedimentary rocks to the west and northwest. The Kansan is a product of Keewatin glaciation. These relationships are common in the type Iowa area and the adjacent Kansan area in northeastern Iowa (Alden and Leighton, 1917, *passim*). In northeastern Goodhue and southeastern Dakota counties, the Iowan overlaps a red drift, whose rocks are Patrician, and is stratigraphically separated from the red drift by a weathering profile. The Patrician drift, in turn, lies upon a Keewatin drift but is not separated from it by a soil or weathering profile. However, it seems reasonable that the Keewatin drift under the Iowan till and loess in southeastern Minnesota and the Keewatin drift separated from the Iowan drift in Dakota County by the Patrician red drift

are the same (Fig. 4). The Patrician red drift remains tentatively classified and correlated with the Illinoian, and the Keewatin drift that underlies the Patrician red drift with the Kansan.

Iowan Drift

Areal distribution.—Leverett (1932, Pl. 2) mapped much of Dakota County as Kansan. This drift was extended southward to the Iowa

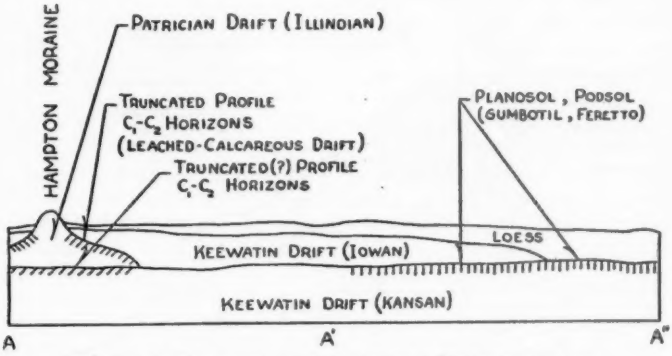

FIGURE 4.—DIAGRAMMATIC STRATIGRAPHIC RELATIONSHIPS OF THE DRIFTS OF SOUTHEASTERN MINNESOTA

WISCONSIN DRIFTS

General Statement

Four Wisconsin drifts have been identified in the area: (1) Iowan, (2) Cary (Keewatin), (3) Cary (Patrician), and (4) Mankato.

Flint (1950) recognized a Cary drift on the James lobe in South Dakota, and Ruhe (1950b) recognized a Cary drift on the Des Moines lobe in southwestern Minnesota and northwestern Iowa. The reason for classifying the Cary drift as such was that the drift occupied the third position in the Wisconsin sequence, and that the third drift of the standard Wisconsin sequence is Cary.

In the Dakota County area stratigraphic, lithologic, and topographic evidences for classification of a Cary (Keewatin) drift are found.

Tazewell drift has not been identified east of the Des Moines lobe in the area.

State line in Fillmore County where the marg. of Leverett's *Kansan* is aligned with the *Iowan* drift border of northeastern Iowa. Leverett discusses the area mapped as Kansan under a heading of *Iowan Drift* (1932, p. 23–39). For many years Leverett considered the Iowan drift a correlative of the late Illinoian (1932, p. 27, 29) and did not accept the Iowan as a separate glacial stage until 11 years after Leighton (1931, p. 51–53; 1933) reclassified it as the earliest substage of the Wisconsin (Leverett, 1942).

The Kansan drift of Leverett has been recognized as Iowan by geologists (Gould, 1941) and soil scientists (McMiller, 1945). This drift, mantled by loess, constitutes the surface drift in this area and is traceable directly southward into the type Iowan (Pl. 1; Fig. 7).

Topography.—In Dakota County the Iowan drift surface is well drained. Closed depressions, characteristic of the Keewatin Cary and Mankato surfaces to the west and the Patrician

PLATE 3.—OLDER TILLS OF DAKOTA COUNTY

FIGURE 1.—ROAD CUT DESCRIBED IN EXPOSURE K-1 OF SECTION 1
K—Kansan; I—Illinoian; Io—Iowan; L—Loess

FIGURE 2.—IOWAN TILL WITH PEBBLE CONCENTRATE AND MANTLE OF LOESS
Sec. 3, T. 113 N., R. 19 W., Dakota County

FIGURE 1

FIGURE 2

OLDER TILLS OF DAKOTA COUNTY

FIGURE 1

FIGURE 2

TOPOGRAPHIC FEATURES OF HAMPTON MORAINE AREA

Cary surface to the north, are lacking. The absence of undrained depressions on the Iowan surface is important for differentiation of this drift from the younger drifts where the two are adjacent. Ruhe (1952a, p. 49–50) has shown that the Iowan is differentiated from the drifts of the Des Moines lobe by erosional topographic discontinuity southward to Hardin County, Iowa.

Slopes in the Iowan area range from 3–6 to 6–12 per cent and relief up to 60–80 feet. All slopes are long and gentle giving rise to an undulating to rolling topography. Drainage lines are integrated.

Lithology.—The Iowan till is light yellow or mottled to gray. Where the Iowan till is overlain by thick loess that is calcareous in its basal part, the Iowan till is calcareous to its top. However, where the overlying loess is thin, both the loess and the upper part of the Iowan till are leached of carbonate. The loess-Iowan till relationships are shown in section 3.

SECTION 3

I-1: NW¼ sec. 27, T.112 N., R19 W., Dakota County

	Thickness Inches
4. Brown to yellowish-brown, well-sorted, leached loess	35
3. Pebble concentrate	4
2. Yellowish-brown, sandy, leached Iowan till	10
1. Yellowish-brown to gray, sandy, calcareous Iowan till	12

I-2: NW¼ sec. 27, T.114 N., R.20 W., Dakota County

	Thickness Feet	*Inches*
4. Yellowish-brown, coarse-textured, well-sorted, leached loess	4	0
3. Yellowish-brown, well-sorted, calcareous loess	5	2
2. Pebble concentrate	0	2
1. Yellowish-brown, plastic, calcareous Iowan till	1	2

Alden and Leighton (1917, p. 156–157) pointed out that where thin loess mantles Iowan till both the loess and upper part of the till are leached. Where thick loess overlies Iowan till, the lower part of the loess is calcareous and is in contact with calcareous till. They reasonably concluded that these stratigraphic relationships proved that loess deposition followed almost immediately the recession of the front of the ice that laid down the Iowan drift.

A pebble concentrate occurs at the top of the Iowan till and under the loess mantle (I-1 and I-2; Fig. 2, Pl. 3). This phenomenon is characteristic also of the "type" Iowan drift (Alden and Leighton, 1917, p. 156).

The Iowan till of Dakota County is a sandy silt. The Iowan texture is comparable to those of the other drifts of Keewatin provenance (Fig. 2). Pebble analyses of several samples are given in Table 1. Composition of the Iowan till is similar to the other drifts of this area that were deposited by glaciers from the northwest.

Iowan stratified drift was not identified in the area. The extensive outwash plains that encircle and are adjacent to the Iowan till may be in part of Iowan age.

The Iowan drift is mantled on the uplands by 18 to 100+ inches of loess. Leverett (1932 Pl. 2) mapped only a very small area in Dakota County as loess. In general, Leverett's loess distribution delineates those areas where loess thickness exceeds 100 inches. The loess-thickness distribution pattern is not understood. In T.114 N., R.20 W., Dakota County, a narrow belt of loess 100+ inches thick trends southeast. A similar belt in T.113 N., R.17 W., when traced southeasterly merges with the main body of the thick loess in Goodhue County. (*See* Leverett, 1932, Pl. 2.) The two local thick belts may have been one continuous deposit. On the present landscape, they are separated by two outwash terraces of Cary and Mankato age. Excavation of broad valleys in the Iowan till plain or resurrection of preglacial valleys by subaerial processes during Tazewell and Tazewell-Cary time may have caused the stripping of thick loess from this local area.

The Iowan age of a part of the loess that mantles the Iowan till in Dakota County is shown by the stratigraphic relationship of

PLATE 4.—TOPOGRAPHIC FEATURES OF HAMPTON MORAINE AREA

FIGURE 1.—ST. PETER SANDSTONE RIDGES TOPPED WITH COARSE ILLINOIAN TILL

FIGURE 2.—TRUE CONSTRUCTIONAL FEATURES OF HAMPTON MORAINE

basal calcareous loess over calcareous Iowan till. Loess deposition followed immediately the uncovering of the till surface by Iowan ice. The time interval was so short that a skeletal soil or weathering profile did not develop in the Iowan till prior to loess deposition. Section I-2 illustrates this relationship.

Apparently increments of younger Wisconsin sediments have been added to the loess of the Iowan area. For example, in the SE¼NE¼ sec. 27, T.112 N., R.19 W., Dakota County, The material above the Iowan till is sandy to a depth of approximately 42 inches. Below this depth and to the top of the Iowan till, the material is loessial (Fig. 5). This section is located a quarter of a mile south of the extensive Cary and Mankato outwash areas in Waterford and Scioto townships. An increment of Cary and/or Mankato eolian sand and silt from a near source has been added to the loess of the Iowan region.

Thus, the composition of the loess in the Iowan area is multiple: (1) part of the loess is of Iowan age, and (2) increments of younger sediments have been added to it. Therefore it is designated as Iowan and post-Iowan loess.[1]

Textures of six typical Iowan and post-Iowan loess samples are shown in Figure 6.

Stratigraphy.—The Iowan drift overlies the Illinoian drift in Dakota County. (*See* K-1 to K-3.) South of the Illinoian drift limits, the Iowan overlies the Kansan drift but is separated from it by a soil or weathering profile in the Kansan till.

The Iowan drift east of the borders of the Keewatin and Patrician Cary drifts is overlain by Iowan and post-Iowan loess.

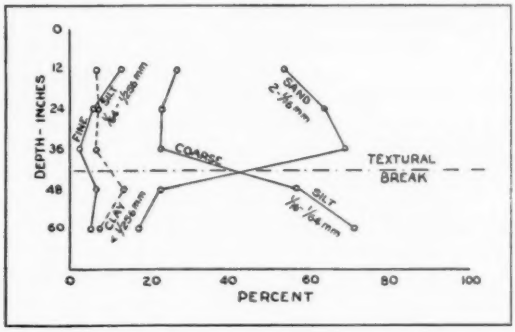

FIGURE 5.—CONTRAST OF TEXTURES OF EOLIAN SEDIMENTS ABOVE IOWAN TILL

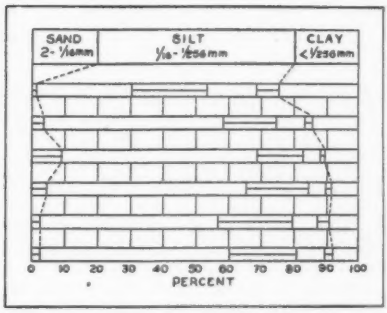

FIGURE 6.—TEXTURES OF TYPICAL IOWAN AND POST-IOWAN LOESSES

Iowan till is exposed under Patrician Cary till in the SC NE¼ sec. 6, T.113 N., R.20 W. (*See* Section 4.)

Cary Drift (Keewatin)

Areal distribution.—The Keewatin Cary drift is here defined as that part of the Des Moines lobe that extends from the margin of the lobe to the margin of the Altamont moraine (Fig. 7; Pl. 1). The location of the Altamont moraine of Leverett (1922) has been modified by Ruhe (1950c; 1952a, *passim*).

Ruhe has traced and mapped this Cary drift on the Des Moines lobe southward through Minnesota to Calhoun and Sac counties, Iowa. Here the Keewatin Cary is the Cary drift

[1] Ruhe (1950c) noted similar characteristics in the loess of the Iowan and younger drift regions in northwestern Iowa and southwestern Minnesota and presented a loess nomenclature that is followed in this report.

mapped by Ruhe (1950b) and correlated through northwestern Iowa, southwestern Minnesota, into eastern South Dakota with the Cary drift of Flint (1950). Thus, regional correlation of the Grantsburg is a correlative of the Altamont moraine (Mankato), whereas the Bemis moraine is overlapped by the St. Croix moraine (Patrician Cary) and is therefore at least Cary in age.

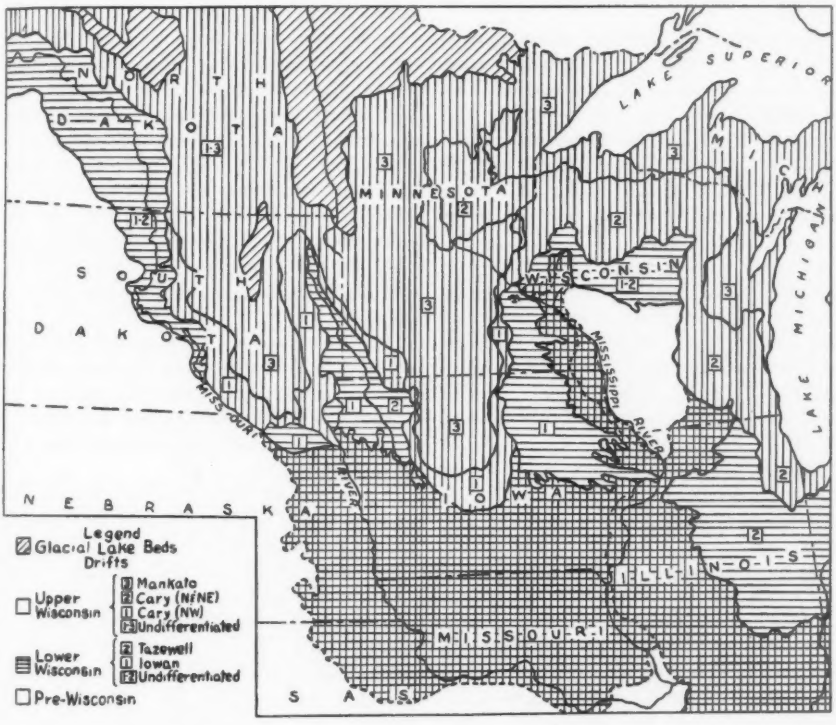

FIGURE 7.—REGIONAL DISTRIBUTION OF THE WISCONSIN GLACIAL DRIFTS

Cary drift of the James lobe in South Dakota and the Des Moines lobe in Minnesota and Iowa is established.

Topography.—The Keewatin Cary till is end morainic. Ridges that trend, in general, north-south are easily identified on aerial photographs and in the field. The end-morainic ridges are part of the Bemis morainic system of Leverett (1932, p. 61). Although Leverett (1932, Pl. 2) mapped a part of the marginal Des Moines lobe as ground moraine, lineation of ridges on the Keewatin Cary surface throughout this area suggests end moraine, and it has been so mapped (Pl. 1). The writers differ from Leverett in correlation of the Grantsburg sublobe of the Des Moines lobe with the Bemis moraine. The Keewatin Cary till surface is poorly drained. Closed depressions are abundant. Integration of drainage on the till surface is in an incipient stage. Ruhe (1952a, p. 49–50) has pointed out that, everywhere around the Des Moines lobe, the Cary drift is differentiated from adjacent older drifts by an erosional topographic discontinuity. Erosional topographic discontinuity is defined as the difference in degree of integration of drainage on the surface of adjacent drifts. The degree of integration of drainage on a drift surface is a reflection of the erosional modification of that surface by subaerial processes after it has been uncovered by and not overridden by a subsequent advance of glacier ice.

In Dakota County, the adjacent older Iowan drift surface is well drained. Integration of surface drainage is complete. Closed depressions are lacking. The expressions of the Keewatin Cary and Iowan topographies reflect their age difference.

Extensive outwash plains and valley trains of Keewatin Cary drift are distributed along Chub Creek, Prairie Creek (northeastern Rice and northwestern Goodhue counties), and Cannon River (Pl. 2). The outwash along Chub Creek heads in the Keewatin Cary end moraine in T.112 N., R.20 W., Dakota County. Along Prairie Creek the outwash is traceable to its head near Cannon City, T.110 N., R.20 W., Rice County, at the margin of the Keewatin Cary end moraine. Both of these outwashes are truncated by the Mankato outwash distributed along Cannon River. Terrace scarps mark the line of truncation of the Chub Creek outwash in T.112 N., R.19 W., Dakota County, and the line of truncation of the Prairie Creek outwash in T.112 N., R.18 W., Goodhue County. North of Lake Byllesby and Cannon Falls, to the north of Cannon River, is a third large plain of Keewatin Cary outwash. The terrace scarp, arcuate and sinuous, extends from the vicinity of Randolph in Dakota County eastward to a point 2 miles east of Cannon Falls in Goodhue County.

The three extensive Keewatin Cary outwash areas constitute the third terrace in the southern part of the area. At the terrace scarps a differential of 10 to 15 feet between the higher Cary plain and the lower Mankato plain is common. The Mankato outwash is the second terrace, which in turn is higher than the post-Mankato alluvial plain along Cannon River (the first terrace; Pl. 2).

Lithology.—Leverett (1932, p. 56–57) described the drift here classified as Cary as a "gray" drift. The color designation may be misleading. Most road cuts expose only the soil profile and the oxidized part of the drift. Colors range from dark gray to gray brown in the soil profile to yellowish brown in the oxidized till. A few deep road cuts expose unoxidized gray till. Particle-size examinations show that, within certain size grades, the sand, which is more than 50 per cent of the constituents, is gray Cretaceous shale fragments. Leverett's term "gray" is applicable to this restricted range of particle sizes.

Analyses of the pebbles of oxidized and unleached Keewatin Cary till show that limestone, basalt, and pink granite are the dominant constituents (Table 1). Shale pebbles constitute only a minor quantity in the analyses. The writers question whether the method of pebble counts accurately shows the composition of a glacial drift. They noted that contrast between two drifts of distinctly different provenances is shown, but that differences between two drifts of similar source may not show up in pebble analyses. Such a relationship is shown in the comparison of the pebble suite of the Keewatin Cary till and the suites of the Kansan, Iowan, and Mankato (Table 1). A Keewatin source is indicated.

Texture of the Keewatin Cary till is a sandy silt (Fig. 2). Similarity of texture of this drift to others derived in the northwest is shown.

Keewatin Cary stratified drift is sandy and gravelly and, at places, calcareous at depths of 25 to 35 inches.

A loesslike silt discontinuously mantles the crests of upland ridges and flats in the Keewatin Cary end moraine and the Keewatin Cary outwash. Ruhe (1950d, *passim*) reported a thin discontinuous mantle of loess on the Cary and Mankato drifts of the Des Noines lobe in northwestern Iowa and southwestern Minnesota.

The loess on the Keewatin Cary drift, in most places, is not more than 24 to 30 inches thick and is patchy. The silt is not found in successive road cuts along a section-line traverse. The loessial silt is brown, and the zone is leached. The till underlying the loess may be calcareous to its top. A calcareous loessial silt has not been observed above a leached Keewatin Cary till or stratified drift. Texture of the loess is variable. Sand content may be as much as 35 to 40 per cent of the total distribution. Four textural analyses of the loessial silt that mantles the Keewatin Cary till are given (Fig. 8, 1–4).

Variability of texture and poor sorting of the loessial silt suggest derivation from near outwash areas or from the Keewatin Cary till surface prior to the encroachment of vegetation. In many places pebbles are concentrated below the loessial silt and at the top of the till. This phenomenon on the Iowan drift has been cited

(Kay, 1931, p. 378) as evidence that the Iowan till was a source for Iowan loess. Deflation of finer constituents resulted in the pebble lag concentrate on the Iowan drift. The same reasoning is applicable to the pebble concentrate on the Keewatin Cary till surface.

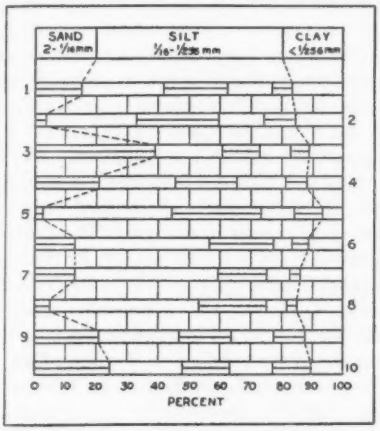

FIGURE 8.—TEXTURES OF UPPER WISCONSIN LOESSES

Stratigraphy.—No exposures were found of Keewatin Cary drift above older Pleistocene deposits. The Cary is differentiated from the adjacent older Iowan drift by a marked erosional topographic discontinuity. Southward in Minnesota, the Keewatin Cary occurs above a weathered profile in Iowan till and above Iowan loess that mantles Iowan drift. In Iowa, many workers have described "Mankato" till above Peorian loess (Kay and Graham, 1943, p. 211–217). The "Mankato" in all these sections has been reclassified as Keewatin Cary (Ruhe, Unpub. rept.). Northward from T.112 N., R.20 W., to T.113 N., R.20 W., the poorly drained end-morainic Keewatin Cary till passes under the Patrician Cary drift approximately at the township line. Several sections (7–10) showing Patrician Cary drift capping Keewatin Cary end-morainic ridges are detailed in this paper.

Cary Drift (Patrician)

Areal distribution.—The Patrician Cary drift is here defined as the Wisconsin red drift of Leverett (1932, p. 39–44) including the St. Croix morainic system and the drift of the of Minneapolis lobe of the Patrician ice sheet of Cooper (1935, p. 6–7) (Pl. 1). The southern limit of the till along the line between Ts.112 and 113 N., R.20 W., is tentative and is based on the southernmost exposures of Patrician Cary till over Keewatin Cary drift. Further study of the Patrician till may result in a southward shift of the border. Leverett (1932, Pl. 2) placed the limit of the Patrician drift several miles north of Lakeville.

The limit of the Patrician till north of the extensive Patrician outwash plain (Pl. 2) does not mark the southern limit of the extension of the Minneapolis lobe of the Patrician ice sheet. A red till outlier occurs in the vicinity of Rich Valley. Two miles south of Coates and near the town of Vermillion, numerous pits, some several hundred feet in longest dimension, occur in the red outwash and attest the presence of Patrician Cary ice in these localities.

Topography.—The Patrician Cary till of the Dakota County area is end morainic. The moraine is typically a knob and kettle complex. (*See* St. Paul quadrangle topographic sheet.) Knobs and end-morainic ridges are roughly aligned and north of the Rosemount plain of Patrician Cary outwash trend parallel to the contract of till-outwash (Pl. 1). In western Dakota County ridges mapped as Patrician Cary trend north-south. Here, however, the Patrician Cary till is thin and discontinuous and mantles the Bemis moraine, Keetwatin Cary, of the Des Moines lobe. Topography is controlled by the Keewatin Cary rather than by the Patrician Cary end moraine.

Relief in the St. Croix moraine in northern Dakota County ranges up to 150 feet. At the end moraine-outwash plain contact, relief on the distal slope of the moraine is only 20–30 feet. The red till outlier at Rich Valley also stands only 20–30 feet above the encircling Rosemount plain. In these areas, the red till is buried under many tens of feet of red outwash.

There is no ground moraine of Patrician Cary drift in the area.

Undrained depressions characterize the stage of development on the till surface. Closed depressions are abundant, and many contain perennial lakes. The poorly drained Patrician Cary till surface compares in degree to that of

the older Keewatin Cary and the younger Keewatin Mankato surface. In topographic expression, these three till surfaces are similar and appear closely related in time.

The poorly drained Patrician Cary till surface contrasts markedly with the well-drained Iowan surface adjacent to it in T.114 N., R.20 W., Dakota County. The erosional topographic discontinuity along the line of contact represents a distinct difference in age between the two surfaces.

TABLE 2.—CORRELATION OF THE TERRACES OF DAKOTA COUNTY

Cannon River set (south)	Vermillion River set (north)
1. Post-Mankato alluvium	1. Post-Mankato alluvium
2. Mankato outwash	2. Mankato outwash
.....	3. Patrician Cary outwash
3. Keewatin Cary outwash	.....

An extensive plain of Patrician Cary outwash is distributed mainly along Vermillion River and its tributaries. The plain is pitted in many places. In the NE cor. sec. 13, T.114 N., R.19 W., 2 miles south of Coates, a pit approximates 250 yards in diameter. Pits are abundant 1–1½ miles north of Vermillion.

The Patrician Cary outwash is the highest terrace (third terrace) in northern Dakota County. This terrace is marked by well-defined scarps where truncated by the lower-level Mankato outwash, which in turn is cut by the post-Mankato alluvium along the present streams. The Mankato terrace is the second terrace, the alluvium the first terrace.

In southern Dakota County, along Cannon River and its tributaries, three terraces also occur. Terraces of the two areas are correlated in Table 2.

The high Patrician Cary outwash terrace is a prominent topographic feature from a point 3 miles north of Farmington to the till-outwash contact near Aligmagnet Lake. The high terrace stands 35–50 feet above the adjacent lower Mankato terrace. The terrace scarp is sharply defined and steep along this traverse.

At Hastings the high terrace stands approximately 250 feet above the Mississippi River Valley fill.

Lithology.—The Patrician Cary till is distinguished from other tills of the area, with the exception of the Illinoian, by its color, texture, and composition. Color ranges from red to reddish brown. The tills of the Keewatin glaciers range from yellowish brown to yellowish gray.

The red to reddish-brown color is primary. In many sections where 25–30 feet of till is exposed, the reddish hue is uniform below the surficial soil. The color is the result of pulverization of ferruginous rock derived from the iron-range regions to the north. The Patrician Cary weathers to brown to gray. In the solum developed in the till, gray-browns are dominant.

The Cary red till is generally sandy to gravelly (Fig. 2). All samples have greater than 50 per cent sand and gravel. Patrician Cary till textures are similar to those of the Illinoian till but markedly dissimilar to the dominantly silty Keewatin tills.

The Patrician Cary till contains an abundance of basalt, granite, quartzite, and sandstone (Table 1). The sandstone is a red variety from outcrops of the clastic rocks of the Lake Superior region. The till is only locally calcareous (Cp-5 to Cp-7, Table 1). The noncalcareous aspect is a primary feature; carbonate has not been leached. Where several tens of feet of Cary till is exposed, the whole section is noncalcareous even though a shallow soil profile caps it. Keewatin tills, however, are invariably calcareous slightly below the capping solum.

In the Rosemount plain, the stratified drift of Patrician Cary age is reddish brown to red, similar to the Patrician Cary till. Plots of till textures (Fig. 2) show a dominance of gravel and sand with silt content of 10–40 per cent and clay generally less than 10 per cent. Analyses of the stratified drift show that the total silt and clay constituted less than 5 percent of the sample.

The stratified drift is poorly sorted. Boulders 10–15 inches in diameter occur in the same zones with coarse sand.

The stratified drift is calcareous. The calcareous nature of the stratified drift is not compatible with the generally noncalcareous nature of its till counterpart. Primary car-

bonate in the stratified drift may be explained by: (1) admixtures of materials from older Keewatin drifts incorporated in the Patrician Cary outwash during the cutting of valleys in the older drifts by meltwaters from the Patrician Cary ice, (2) contamination of the Patrician Cary stratified drift by Paleozoic limestone as the meltwaters either scoured valleys or occupied valleys in the Paleozoic rocks, and/or (3) the contemporaneous deposition and mixing of stratified drifts from the Keewatin Cary and Patrician Cary ice masses.

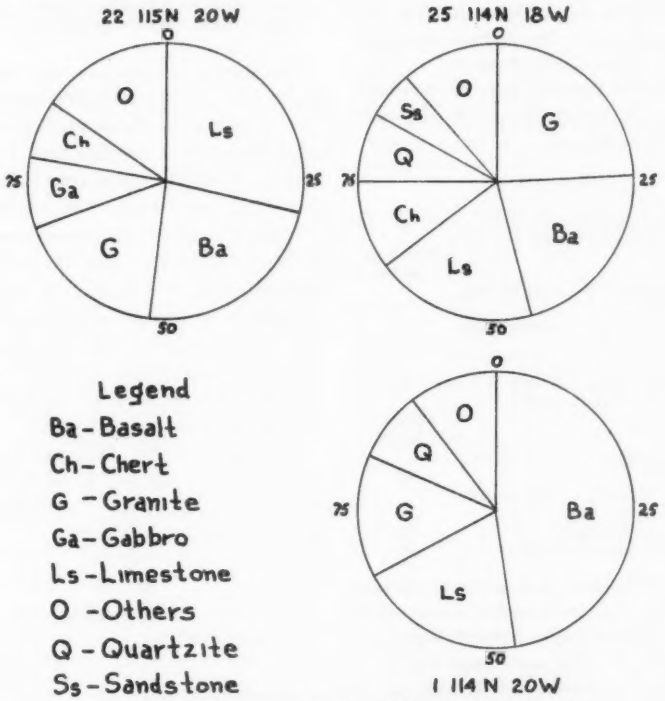

FIGURE 9.—COMPOSITION OF CARY (PATRICIAN) OUTWASH

(1) The Patrician Cary outwash occupies a terrace position in a broad valley cut in the Keewatin Iowan and older drifts. (*See* Schwartz, 1936, p. 70–72 for glacial and preglacial excavation hypotheses.) Although the broad valley may have been excavated during Tazewell, Tazewell-Cary, and Keewatin Cary time,[2] during Patrician Cary time meltwater from the Patrician ice may have modified, cut, and incorporated older drift materials with resultant contamination and increase in carbonate content.

(2) Outcrops of Paleozoic limestones may have been available for scour and incorporation by meltwaters of the Patrician Cary ice, or deposition of a Patrician till mantle may have prevented access to the Paleozoic limestones.

(3) The physiographic similarity of the Keewatin Cary and the Patrician Cary till surfaces suggests a close time relationship for

[2] Occurrence of Iowan drift in Washington County, Minnesota, Pierce County, Wisconsin, and near Red Wing in Goodhue County, Minnesota, and, in addition, the occurrence of Iowan till inliers in the Patrician Cary outwash suggests: (1) Iowan ice completely covered the broad valley, (2) Iowan till was deposited in the valley to elevations approximately accordant with the Iowan upland south of the outwash, (3) sub-aerial excavation occurred during the glacial and intraglacial intervals cited, and (4) that an original pre-glacial control is not precluded.

these two glacial episodes. The outwash terrace relationships in the Dakota County area are Keewatin Cary and Patrician Cary outwashes truncated by Mankato, but in no place is Keewatin Cary outwash truncated by Patrician Cary. This suggests that the Patrician Cary outwash may be in part of Keewatin Cary age with the admixture of the primary carbonate characteristic of the Keewatin drift.

A loesslike silt mantles both the Patrician Cary till and outwash. Section 4 illustrates loess-till relationships.

SECTION 4

Cp-1: NW¼ SW¼ sec. 20, T.115 N., R.20 W., Dakota County (*See* Fig. 8, no. 5)

	Thickness Inches
2. Gray, noncalcareous, well-sorted, loesslike silt	24
1. Red, noncalcareous, cobbly till, exposed	36

Cp-2: NC sec. 18, T.114 N., R.20 W., Dakota County (*See* Fig. 8, no. 6)

	Thickness Inches
4. Yellowish-brown, well-sorted, leached loess	44
3. Yellowish-brown, well-sorted, calcareous loess	22
2. Red, cobbly, leached till	62
1. Red, cobbly, calcareous till	40

In Cp-2, the leached till crops out at the crest of a small hill. The loess mantles the east slope of the hill; the slope is approximately 8 per cent. On the slope, the calcareous basal loess is in contact with calcareous, cobbly red till. A weathering profile does not occur between the loess and till.

The loesslike silt-Patrician Cary outwash relationships are shown in section 5.

SECTION 5

Cp-3: SC SE¼ sec. 23, T.115 N., R.20 W., Dakota County (*See* Fig. 8, no. 7)

	Thickness Feet	Thickness Inches
2. Yellowish-brown, moderately well sorted, leached loess	3	2
1. Red, bouldery, calcareous outwash	15–20	

Cp-4: NW¼ sec. 9, T.114 N., R.19 W., Dakota County (*See* Fig. 8, no. 8)

	Thickness Feet	Thickness Inches
2. Yellowish-brown, well-sorted, leached loess	3	1
1. Red, bouldery, calcareous outwash; red sandstone and basalt boulders abundant	5	5

Cp-5: SC NE¼ sec. 25, T.114N., R.18W., Dakota County (*See* Fig. 8, no. 9)

	Thickness Feet	Thickness Inches
3. Reddish-brown, poorly sorted, leached, loesslike silt	1	3
2. Red, leached outwash; abundant red sandstone and basalt boulders	1	5
1. Red, calcareous outwash; abundant limestone and dolomite pebbles	2	8

The section descriptions and the textural analyses lead to several generalizations on the loess mantling the Patrician Cary drift. The loess ranges from 2 to 4 feet thick. Color varies from a yellowish brown to gray to brown. Texture of the loess ranges from poorly sorted to well sorted. It may be calcareous, noncalcareous or leached depending upon the availability of carbonate and modification of the loess by weathering subsequent to deposition. Patrician Cary age is proved by the stratigraphic relationship of basal calcareous loess over calcareous Patrician red till with no weathering profile between the two (section Cp-2).

Stratigraphy.—The Patrician Cary drift mantles Iowan drift and Keewatin Cary drift. The relationship of the Cary red till to the Iowan is shown in section 6.

SECTION 6

Cp-6: SC NE¼ sec. 6, T.113 N., R.20 W., Dakota County (road cut)

	Thickness Feet	Thickness Inches
2. Patrician Cary till		
Gray to yellowish-brown, noncalcareous, sandy till	1	6
Red, noncalcareous till; abundant red sandstone and basalt pebbles	6	6
1. Iowan till		
Yellowish-brown, compact, plastic, leached till; 22 inches exposed in cut, 40 inches in boring	5	2
Yellowish-brown, compact, plastic, calcareous till; primary carbonate; 8 inches in boring	0	8

In section 6, the marked contrast in color, composition, and texture differentiates these tills both physically and in regard to provenance. The upper till is a Patrician type, and the lower is a Keewatin type. The weathering profile in the lower till differentiates the two tills in relative time such that the lower is earlier Wisconsin.

In several sections the Patrician Cary drift mantles Keewatin Cary drift, the marginal part of the Des Moines lobe (Section 7).

SECTION 7

CP-7: SW cor. sec. 33, T.113 N., R.20 W., Dakota County (road cut)

	Thickness Feet
2. Red, noncalcareous gravel; abundant basalt and red sandstone cobbles (Patrician Cary)	2
1. Yellowish-brown, calcareous till; abundant granite, basalt, limestone pebbles, Cretaceous shale moderately abundant (Keewatin Cary)	15–20

The upper drift veneers a gentle southern slope (2–3 per cent) of a morainic ridge that trends north-south. The topographic expression of the morainic ridge is controlled by the lower yellowish-brown till. The morainic topography trends north-south throughout this area, and its margin is 2 miles east of section 7 along the Minneapolis, Northern, and Southern Railway. Leverett (1932, Pl. 2) mapped this area as the Bemis moraine of the young gray Wisconsin drift. The lower till of section 7 has been traced southward by Ruhe to Des Moines, Iowa. This till is the marginal part of the Des Moines lobe.

Lithology of the upper drift is given in Table 1 (Cp-1). The rock types are characteristic of the Lake Superior region. A Patrician source of this drift is evident. The lower till (Table 1, Ck-1) contains a different suite of rock types. This lithology is characteristic of areas to the west and northwest and indicates a Keewatin source.

Section 7 proves that the marginal part of the Des Moines lobe, Keewatin Cary of the writers, is overlapped by a Patrician drift that has been recognized as Cary (Glacial Map of North America, 1945). The margin of the Des Moines lobe must be removed from the Mankato and placed as pre-Patrician Cary.

In section 7 a soil or weathering profile does not separate the two drifts. It has been noted that the constructional morainic topographies of these drifts are similar. Therefore the lower drift is believed to be closely related in time to the previously recognized Cary of Patrician provenance. The lower drift therefore has been classified as Cary but of Keewatin provenance.

Section 8 shows similar stratigraphic relationships.

SECTION 8

Cp-8: SC NW¼ sec. 6, T.113 N., R.20 W., Dakota County

	Thickness Feet	*Inches*
2. Red, noncalcareous till; abundant red sandstone, basalt, granite cobbles (Patrician Cary)	2	6
1. Yellowish-brown, calcareous till; abundant limestone and granite pebbles (Keewatin Cary)	5	0

Lithologies of these tills are given in Table 1 (Cp-2, the upper till; CK-2, the lower till). The upper till is a thin patchy mantle on a north-south trending morainic ridge of the lower till. The similarity of sections 7 and 8 is evident.

Sections 9 and 10 near Eureka Center, Dakota County, offer conclusive proof of the overlap of the marginal Des Moines lobe by the Patrician Cary drift.

SECTION 9

Cp-9: EC sec. 21, T.113 N., R.20 W., Dakota County (gravel pit)

	Thickness Feet	*Inches*
2. Keewatin Cary till		
Gray, sandy to silty, leached till	1	8
Gray, sandy, calcareous till; limestone pebbles abundant	1	4
1. Keewatin Cary stratified drift; bedded, sorted, calcareous bedding truncated to westward by base of till	15–20	

The gravel pit is opened in a flat-crested hill approximately 200–250 feet in diameter. Lithology and the moderate weathering on an upland flat position indicate that the drift is of the Des Moines lobe. Leverett (1932, Pl. 2) mapped it as young gray drift. The stratified drift in the section may be the remnant of a proglacial delta subsequently overridden by the Keewatin Cary ice with resultant till deposition.

A quarter of a mile east of the gravel pit, in cuts along the Minneapolis, Northern, and Southern Railway and an adjacent road, is section 10.

SECTION 10

Cp-10: WC NW¼ sec. 22, T.113 N., R.20 W., Dakota County

	Thickness Feet
2. Red, noncalcareous drift; abundant basalt and red sandstone boulders up to 12 inches in diameter; 1 foot thick at crest; thickens to 8 feet on flanks (Patrician Cary)	1–8
1. Gray to yellowish-brown, calcareous drift; abundant limestone pebbles (Keewatin Cary)	
Sandy to silty calcareous till	4
Bedded, calcareous stratified drift	11

No. (1) in Section 10 is the same horizon as the till and stratified drift exposed in the gravel pit to the west (Section 9), but in Section 10 the Keewatin Cary drift is mantled by a thin bouldery Patrician Cary drift.

Throughout the area of the section descriptions (7–10), the Keetwatin Cary is the dominant drift. It is mantled in places by thin, small patches of Patrician Cary drift. Stratigraphic evidence proves that the marginal part of the Des Moines lobe, previously considered Mankato, is overlapped by the Cary drift of Patrician provenance and is therefore pre-Patrician Cary. The writers consider this drift Cary of Keewatin provenance.

Mankato Drift

Areal distribution.—The Mankato drift is here defined as that part of the Des Moines lobe that is margined by the Altamont moraine. The Altamont moraine is used as defined by Leverett (1922, p. 153), but not as described in older literature. The location of the Altamont moraine of the Des Moines lobe has been modified by Ruhe (1950c; 1952a, *passim*).

Areal distribution of the Mankato drift is shown in Plate 1, and its regional distribution is shown in Figure 7. The Mankato drift has been traced and mapped on the Des Moines lobe by Ruhe (Fig. 7) southward through Minnesota into Calhoun and Sac counties, Iowa, where the margin of the Mankato drift joins the margin of the Mankato drift mapped previously by Ruhe (1950b). The Mankato drift of the latter area was correlated through northwestern Iowa and southwestern Minnesota with the Mankato drift of Flint (1950) in eastern South Dakota. Thus regional correlation of the Mankato drift of the James lobe in South Dakota and the Des Moines lobe in Minnesota and Iowa is established.

Topography.—The Mankato till surface is mainly end morainic. Ridges, in general, parallel the margin of the Mankato drift and trend north-south. In the area between the Minnesota and Mississippi rivers, trends change to the northeast. This area is the juncture of the Altamont morainic system of the Des Moines lobe and the Grantsburg sublobe.

Leverett (1922, p. 61) stated: "The Rush lake morainic system of the Grantsburg sublobe, with its four members, seems to be the equivalent of the Bemis morainic belt." The writers present the following evidence that the marginal moraine of the Grantsburg sublobe is the equivalent of the Altamont moraine, the marginal moraine of the Mankato drift: (1) Stratigraphic evidence has been detailed to prove overlap of the margin of the Des Moines lobe, the Bemis moraine of Leverett, by the Patrician Cary drift. (*See* Leverett, 1932, p. 61 for discussion of the Bemis moraine in Rice and Dakota counties, Minnesota.) (2) Morainic ridges in the Altamont moraine in western Dakota County and eastern Scott County trend generally north-south. At the base of the Grantsburg sublobe, trends change to northeasterly and parallel the margin of the gray drift. The Patrician Cary drift in northern Dakota County trends, in general, east-west. In northwestern Dakota County an angular topographic discontinuity (Ruhe, 1952a, p. 52–54) approximating a right angle separates the trends of the Altamont moraine (Mankato) from those of the St. Croix moraine (Patrician Cary) (Pl. 1). (3) Stratigraphic evidence (Sections 12, 13) shows the overlap of the Patrician Cary drift by the Mankato drift of the Altamont moraine. Thus, the Grantsburg sublobe is not correlative of the drift of the Bemis moraine but is the equivalent of the Altamont moraine. The Grantsburg sublobe is Mankato.

The Mankato till surface is poorly drained and is similar in its stage of drainage development to the Patrician Cary and the Keewatin Cary surfaces. Ruhe (1950a, p. 439–442), on the basis of the frequency distributions of slopes on

the poorly drained Cary (Keewatin) and Mankato surfaces in northwestern Iowa, concluded that these two drifts were very closely related in time. The same conclusion is applicable in the Dakota County area except that the two drifts are separated stratigraphically by the Patrician Cary drift. Hence in this area there are three late Wisconsin drifts closely related in time.

Areal distribution of the Mankato outwash terrace is shown in Plate 2. Topographic relationships of the Mankato terrace to the Keewatin Cary and Patrician Cary outwash terraces have been described; correlation of the terraces is given in Table 2.

Lithology.—The Mankato till has physical characteristics similar to those of the Keewatin Cary till. Color varies from yellowish brown in the oxidized zone to gray in the unoxidized zone. Leverett (1932, p. 56–75) considered this drift, as well as the Keewatin Cary, his Wisconsin gray drift (Des Moines lobe). The designation "gray" applied to the Mankato is as misleading as when applied to the Keewatin Cary. "Gray" is applicable only to the unoxidized till which is not exposed at many places in the area.

Plots of textural analyses of typical samples of Mankato till (Fig. 2) fall within the textural range of tills derived from northwest of the area—*i.e.*, of Keewatin provenance. Mankato till, in general, is a sandy silt.

Composition of the Mankato till is similar to the Keewatin Cary. Pebble suites showed dominance of limestone, granite, and basalt. Cretaceous shale particles constitute a significant part of the assemblages. Mankato till composition differs distinctly from the Patrician Cary (Table 1).

Texture of Mankato stratified drift is sandy to gravelly. In places, the gravel contains boulders up to 10 inches in diameter. Composition is similar to that of the Mankato till. Granite, basalt, and limestone particles constitute the bulk of the material.

A thin loesslike silt, patchy in areal distribution, overlies the Mankato till and stratified drift. Ruhe (1950d, *passim*) reported a similar loess on the Mankato drift in southwestern Minnesota and northwestern Iowa.

Section 11 is typical of Mankato loess and its relationship to the Mankato till.

SECTION 11

M-1: NC NE¼ sec. 9, T.111 N., R.22 W., Rice County, Minnesota. (*See* Fig. 8, no. 10)

	Thickness Inches
3. Brown, leached, sandy, loesslike silt	27
2. Yellowish-brown, leached, pebbly till	21
1. Yellowish-brown, calcareous, pebbly till	14

The abundance of sand (24 per cent) suggests a near source for the loess, perhaps Mankato outwash or the adjacent till surface, prior to the encroachment of vegetation.

Relationship of Mankato loess to Mankato outwash is shown in section 12.

SECTION 12

M-2: SW cor. sec. 24, T.114 N., R.19 W., Dakota County

	Thickness Inches
3. Dark gray to brown, leached, sandy, poorly sorted, loesslike silt	22
2. Yellowish brown, leached, poorly sorted, loesslike silt	12
1. Yellowish brown, leached, cobbly, sand and gravel; abundant granite and basalt cobbles	27

Section 12 is located on the Farmington plain of sand and gravel that can be traced to the margin of the Altamont moraine in the western part of Dakota County. The higher Rosemount plain (Patrician Cary outwash) is truncated by the Farmington plain three quarters of a mile north and 1 mile south of this section.

No weathering or soil profile was observed between the thin, patchy Mankato loess and the Mankato till or outwash. Leached loess overlies either leached or calcareous till or outwash. Calcareous Mankato loess was not observed over leached Mankato till or outwash. These stratigraphic relationships suggest that loess deposition occurred with deglaciation, and that the sediments, till, outwash, and loess are parts of a single glacial episode.

Stratigraphy.—Section 13 shows the stratigraphic superposition of Mankato till on Patrician Cary red drift.

SECTION 13

M-3: SE cor. sec. 23, T.115 N., R.21 W., Dakota County (road cut at intersection of U.S. Highway 65 and County Road 11)

	Thickness Inches
2. Mankato till	
Leached, yellowish-brown to gray, silty till	39
Calcareous, yellowish-brown, silty till; abundant limestone pebbles	14
1. Patrician Cary drift	
Red, calcareous drift; abundant basalt and red sandstone pebbles, no limestone pebbles, carbonate may be secondary; exposed	18

M-4: SC sec. 29, T.115 N., R.20 W., Dakota County (road cut)

	Thickness Inches
2. Mankato till	
Leached, yellowish-brown, silty till	39
Calcareous, yellowish-brown, silty till, with limestone pebbles	12
1. Patrician Cary drift	
Red, calcareous, gravelly drift; abundant basalt and red sandstone pebbles; no limestone pebbles, carbonate may be secondary	36

M-5: SC SE¼ sec. 1, T.27 N., R.23 W., Dakota County (road cut)

	Thickness Feet
2. Mankato till: yellowish brown to gray, calcareous; occurs as thin smear or morainic knob	1–2
1. Patrician Cary till: red, non-calcareous; abundant basalt and red sandstone pebbles	8–10

The exposures of Section 13 are in morainic topography near the margin of the Mankato drift. In each the Mankato till veneers the underlying Patrician Cary drift. In all cases the basal Mankato till is calcareous and may overlie either a calcareous or noncalcareous Patrician Cary drift. Evidence shows that the noncalcareous nature of the Patrician Cary is primary and not the result of weathering. Thus, an intraglacial interval of weathering or soil development is not represented by these stratigraphic relationships.

Two interesting exposures (Section 14) in southeastern Scott County show Mankato till over Patrician Cary till.

SECTION 14

M-6: SE¼, sec. 36, T.113 N., R.21 W., Scott County (road cut)

	Thickness Feet
2. Mankato till: yellowish-brown to gray, calcareous till with abundant granite, basalt, and limestone pebbles and cobbles	8
1. Patrician Cary till: red, calcareous till with abundant red sandstone, basalt, granite, and limestone pebbles	12

M-7: SE¼ SW¼ sec. 35, T.113 N., R. 21 W., Scott County (road cut)

	Thickness Feet
2. Mankato till: yellowish-brown to gray, calcareous till with abundant granite, basalt, limestone pebbles and cobbles	15
1. Patrician Cary till: slightly calcareous, red till with abundant basalt and red sandstone pebbles	8

No soil or weathering profile separates the two drifts. Section 14 is part of a series along Castle Rock road that shows the relationships of the Keewatin Cary, the Patrician Cary, and the Mankato drifts. Along the township line road 2 miles west of Castle Rock, the border of the Keewatin Cary drift (margin of the Des Moines lobe) is crossed. Two miles west of the border is the section of Patrician Cary drift over Keewatin Cary till (Section 7). A mile and a quarter west of this section, the subdued morainic margin of the Mankato drift (Altamont moraine) is crossed. The sections of Mankato till over Patrician Cary till (Section 14) are located 1 mile and 2¼ miles west of this drift border. Thus, along the township line traverse of 5½ miles, topographic, lithologic, and stratigraphic relationships of the three upper Wisconsin drifts are discernible.

Classification of the Wisconsin Glacial Stage

The generally accepted classification of the Wisconsin glacial stage (Leighton, 1933, p. 1681) is a fourfold division:

Wisconsin { Mankato, Cary, Tazewell, Iowan

Ruhe (1952b, *passim*) objected recently to this classification mainly on the basis that all

intraglacial[3] intervals in the sequence were recognized as equivalent in magnitude. Ruhe detailed evidences (p. 398–401) to show that the major break within the Wisconsin glacial stage is the Tazewell-Cary interval, and that this interval is the correlative of the time of Brady soil genesis of the loesses of Kansas and Nebraska. Previously, the Brady soil had been correlated with the Two Creeks Forest Bed, Cary-Mankato interval (Schultz and Stout, 1945, p. 243; 1948, p. 573). Ruhe suggested the following classification:

Wisconsin Glacial Stage	Upper Wisconsin	Mankato
		Cary
	Bradyan interval	
	Lower Wisconsin	Tazewell
		Iowan

In the Dakota County area, three drifts in the Upper Wisconsin complicate the problem of classification: Cary (Keewatin), Cary (Patrician), and Mankato. The Cary (Keewatin) is the correlative of the Cary drift of the James lobe in South Dakota (Flint, 1950) and the Cary drift of the Des Moines lobe in Iowa and southwestern Minnesota (Ruhe, 1952b, p. 399–400). The Cary (Patrician) is the correlative of the Cary of the original classification of the Wisconsin glacial stage (Leighton, 1933, p. 168; Glacial Map of North America, 1945). Thus, one aspect of the dilemma of the classification is to place three glacial drifts, each distinctive and equivalent in rank to the other two, into two classification categories in order to maintain harmony with previous classifications. If a classification had not existed, each drift probably would have been named, and each drift recognized as the equivalent of a substage.

In the Cary drifts, the terms "Keewatin" and "Patrician" have been used, not to connote deposition by independent glaciers, but in the sense of provenance. Flint (1943, p. 330) has raised objection to the terms "Keewatin" and "Patrician" as follows:

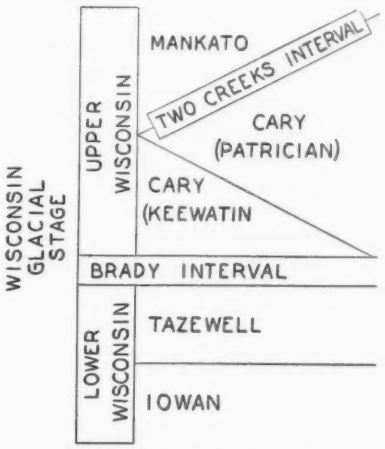

FIGURE 10.—PROPOSED CLASSIFICATION OF THE WISCONSIN GLACIAL STAGE

"It is unfortunate that drift sheets in the Mississippi basin have been identified, on the basis of lithology, as Labradorian drift, Patrician drift, and Keewatin drift, whereas all that is proved by their lithologic characters is that the ice responsible for them traversed, respectively, regions southeast, south, and southwest of Hudson Bay."

Flint recognizes provenance in this quotation. The writers believe that most students of geology are familiar with the general locations of the "centers" Keewatin and Patrician, and that the terms are useful in the sense of provenance; hence they have been applied to the Cary drifts. Physical characteristics of the drifts are too variable regionally to be used in a naming sense, although locally they may be quite constant.

Flint (1947, p. 251) has emphasized the magnitude of the Two Creeks Forest Bed (Two Creeks Interval) in the Wisconsin sequence. The Two Creeks interval occurs between the Mankato and Cary (Patrician) of the present

[3] "Intraglacial" is here used in the sense of Kay and Graham (1943, p. 156).

classification. In the Dakota County area it is not represented. The stratigraphic break between the Mankato and Cary (Patrician) as well as between the Cary (Patrician) and Cary (Keewatin) are very minor. No weathering or soil profile was found within the Upper Wisconsin.

The writers believe that to minimize confusion in regard to the classification of the Wisconsin glacial stage the terms "Keewatin" and "Patrician" may be logically applied in a provenance sense to the Cary drifts of the Dakota County area (Fig. 10).

The writers predict that as glacial studies progress, and as detailed studies are undertaken farther north, other complications in classification will arise and that division of now recognized substages will be necessary.

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MISSISSIPPI RIVER AND GULF COAST TERRACES AND SEDIMENTS AS RELATED TO PLEISTOCENE HISTORY—A PROBLEM

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ABSTRACT

Direct evidence of low and high sea levels accompanying Pleistocene glacial and interglacial Stages is being sought in marine terraces on the East and West Gulf coast, in fluvial terraces and deltaic plains in the valley of the Mississippi River, and in sediments of the northern Gulf Coastal Area and adjacent Continental Shelf.

Marine terraces on land in Georgia, Florida, Alabama, and Mississippi suggest interglacial stages when sea level was high. The shorelines are generally considered to range from the highest and oldest to the lowest and youngest. A net uplift of the land of the East Gulf Coastal Plain relative to sea level during the Pleistocene is indicated. This is thought to have been caused by tectonic forces rather than by progressively decreasing degrees of deglaciation. However, much difference of opinion exists concerning the number and elevations of old shore lines and especially concerning their ages from Pliocene to Thermal Maximum.

Shore features are notably lacking on land along the Louisiana and Texas coasts. Pleistocene subsidence on this part of the Gulf Coast is evident.

Four alluvial terraces and recent alluvium in the lower Mississippi Valley are thought by Fisk and Russell to record stream deposition in Aftonian, Yarmouth, Sangamon, Brady, and Post-Mankato times, and deep erosion during intervening glacial intervals. However, Horberg, Leighton, and Willman consider that three of these four terraces are pre-Pleistocene and that the fourth is a valley train deposited during the Mankato glacial Subage.

Sedimentary formations on the Gulf Coast give some evidence of repeated emergence and erosion followed by submergence and deposition, but again the number of formations and disconformities and their Pleistocene ages are uncertain.

Known occurrences of sand and gravel and of bioherms in deep water near the outer edge of the Continental Shelf suggest one or more low stands of sea level, but their ages are unknown.

It seems doubtful that any part of the Mississippi Valley and Gulf Coast area remained stable tectonically during the Pleistocene.

Future complete cores from present sedimentary surface to Tertiary on coastal land or tideland may yield a complete record of Pleistocene history.

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Pleistocene Classification

The standard classification of the Pleistocene for North America is shown in the left-hand columns of Figure 1. The pre-Wisconsin glacial and interglacial Stages are of long standing. The detailed divisions of the Wisconsin have been recognized more recently but are now well established. The Farmdale, Iowan, and Tazewell glacial Substages, or Subsubstages of the Early Wisconsin, and the Cary and Mankato Subsubstages of the Late Wisconsin are so nearly of the same age that for present purposes the Early Wisconsin, Mid-Wisconsin (Brady), and Late Wisconsin divisions are sufficient. The Post-Mankato and its subdivisions, however, should not be omitted for these are parts of the Wisconsin and of the Pleistocene. The Wisconsin and the Pleistocene have not ended, and the Recent is eliminated. The Cochrane, which represents a readvance or at least a halt of an ice sheet in Manitoba, Ontario, and Quebec, is here placed below the Thermal Maximum in agreement with Antevs (1953, Fig. 1, p. 197, 206), rather than above the Thermal Maximum as suggested by Flint (1952; 1953, p. 914). At its climax the southern border of the Cochrane ice sheet was 575 miles from its center of accumulation and 675 miles from the maximum extent of the Mankato glacier (Antevs, 1953, p. 204). This large ice sheet could hardly have melted away at the same time that the less extensive glaciers of the Little Ice Age were forming.

Pleistocene Chronology

In column 2 of Figure 1 estimates of the durations of the older interglacial Ages, the Mid-Wisconsin Subage, and the four Post-Mankato subdivisions are given. In general these figures are determined by averaging previously published estimates and modifying the averages in the light of recent determinations based on radioactive carbon, varve counts, pollen analysis, and rates of retreat. No claim is made to accuracy in details.

The duration assigned to the Mid-Wisconsin interval may seem out of proportion. By evidence in Iowa (Ruhe, 1952a; 1952b), Nebraska (Schultz and Stout, 1945; Thorp, Johnson, and Reed, 1951), and Kansas (Frye and Leonard, 1952), the writer is convinced that the Brady interval was much longer than Post-Mankato time. In making it only twice as long the writer is being conservative.

The amount of time that has elapsed since the maximum stand of the last of the great Pleistocene ice sheets has been variously estimated. Kay's figure (G. F. Kay, 1931) often quoted is 25,000 years. Flint (1947, p. 406, 532) made a rough estimate of 13,500 years. Determinations by the radiocarbon method average about 11,000 years (Flint and Deevey, 1951, p. 263). Antevs (1953, p. 197) places it at 20,000 years. About 15,000 years seems a reasonable figure (Fig. 1).

There is considerable difference of opinion concerning the dating of the Thermal Maximum and the Little Ice Age. It has generally been thought that the Thermal Maximum started about 6000 years ago and ended about 4000 years ago. Antevs (1953, p. 197) places it from 7000 to 4500 years ago. Ahlmann (1953, p. 38), however, concludes that the Thermal Maximum (Climatic Optimum to him) was about 6000 years long and occupied the time

between 9000 and 3000 years ago. According to a rough average of these figures the Thermal Maximum started about 7500 years ago, lasted about 3500 years, and ended about 4000 years ago. The main points are that the climate was warmer than it is now for some thousands of years, causing notable reduction of previously existing land ice, and that about 4000 years ago a glacial rejuvenation was started that resulted in the glaciers of the Little Ice Age. Records show that existing glaciers were at the maximum as late as 200 years ago and that since then they have become somewhat smaller and thinner.

CLASSIFICATION			ESTIMATED DURATION (YEARS)	THEORETICAL SEA-LEVEL VARIATIONS	
WISCONSIN	Post-Mankato	Present	200	0	Sea level rising slowly and slightly glaciers shrinking
		Little Ice Age	3800	-2(?)	Slightly more ice than now
		Thermal Maximum	3500	+30	20% less ice than now
		Cochrane	7500	-30	20% more ice than now
	Late Wisconsin	Mankato Cary-Mankato (Two Creeks) Cary		-300	Late Wisconsin maximum
	Mid-Wisconsin	Tazewell-Cary (Brady)	30,000	+75	50% less ice than now
	Early Wisconsin	Tazewell Iowan-Tazewell Iowan Farmdale-Iowan Farmdale		-300	About same ice volume as late Wisconsin
Sangamon			70,000	+150	All ice melted
Illinoian				-450	Ice volume somewhat less than Pleistocene maximum
Yarmouth			180,000	+150	All ice melted
Kansan				-450	About same as Illinoian
Aftonian			120,000	+150	All ice melted
Nebraskan				-450	About same as Illinoian and Kansan

FIGURE 1.—PLEISTOCENE CLASSIFICATION, DURATIONS OF SIGNIFICANT TIME UNITS, AND THEORETICAL SEA-LEVEL VARIATIONS

THEORETICAL SEA-LEVEL VARIATIONS

A reasonable average of 8–11 estimates previously made suggests that, if all the glacier ice now existing on the lands of the world should melt, sea level would be raised about 120 feet; that, during the Cary-Mankato maximum, sea level would have been about 240 feet lower than it is now; and that, during the maximum of all Pleistocene glaciation, sea level should have stood at about −380 feet. According to Ahlmann (1953, p. 3), however, the Greenland and Antarctican ice sheets are now known to be thicker than was formerly supposed and, if all existing ice were melted, sea level would rise at least 198 feet. On the assumption that Ahlmann may be more nearly right than others, it is concluded that the melting of all existing land ice might raise sea level 150 feet, an increase of 25 per cent over the average of previous estimates. The same percentage increase in volume for the older Pleistocene ice sheets would indicate a 300-foot lowering of sea level at the Late Wisconsin Climax and 475 feet at the maximum of all Pleistocene glaciation.

By taking some such figures as a guide, the theoretical sea level for each of the glacial and interglacial intervals that are important to this study can be estimated (Fig. 1). The Scandinavians who first presented evidence for the Thermal Maximum (Climatic Optimum) would probably raise sea level for this interval, for they believe or at least some of them believe that almost all land ice may have been melted at this time. The Illinoian, Kansan, and Nebraskan ice sheets were nearly the same size,

and each was smaller than the maximum Pleistocene estimates that included all three.

These are the roughest sort of estimates and should not be taken too seriously.

Marine Terraces as Records of Sea-Level Changes

Preliminary Statement

If sea level varied as indicated, there should be a record, at least of the interglacial intervals, in old shore lines now above sea level. Shore lines formed during glacial ages when sea level was lower than it is now would be submerged and not available. Raised shore lines that would match the interglacial intervals would not be expected in the coastal areas of glaciated regions, for there the land surfaces subside by loading during glaciation and rise when unloaded in the interglacial intervals—just the opposite of this discussion. Neither should the subject shore lines by looked for in nonglaciated areas that were subject to ordinary diastrophic movements—uplift, subsidence, folding, warping, tilting, faulting—during the Pleistocene. If such movements had taken place, the raised shore lines would not match the interglacial intervals considered. If a given land mass were elevated faster and farther than sea level rose in the early stages of an interglacial interval, or if the sea floor anywhere subsided greatly, the resutling interglacial shore line might be minus rather than plus. Similarly, but in reverse, land might be submergent rather than emergent during a glacial age. Raised shore lines should be sought on the borders of unglaciated lands that were diastrophically stable during the Pleistocene, if such lands exist. These shore lines should exist in all continents and islands throughout the world. A shore line at a given level might be occupied during more than one interglacial Age if the degree of deglaciation were the same and there were no diastrophic movement of the land or the sea floor.

Aftonian or Yarmouth shore lines may have been eroded past recognition by now. For each raised shore line there should be marine deposits at the same and lower levels. Such deposits might remain after the topographic shore lines had been eroded. Or such shore lines may be buried under younger nonmarine deposits.

In North America, the most likely places for this record of eustatic variation in sea level due to deglaciation are the southern part of the Atlantic Coastal Plain, and the Coast of the Gulf of Mexico.

Atlantic and East Gulf Coasts

Cooke (1931; 1935; 1945, p. 12–13, 17, 245–248, 273–314; Parker and Cooke, 1944, p. 20–27) presented, named, and described eight shore lines and formations on the Atlantic and East Gulf coasts, complete with elevations, Pleistocene dates, and origins (Fig. 2). For Cooke, Post-Mankato history was the simple melting of the Mankato ice and the resulting rise of sea level from far below present sea level to its present position. If he considered the effects of the Cochrane, Thermal Maximum, and Little Ice Age episodes, he evidently thought them of minor importance and considered only net retreat. With the disappearance of the continental glacier the sea had reached its present level and has "remained almost constant ever since" (Cooke, 1945, p. 313). The implication is that all the Mankato ice melted except that existing today, and that then melting stopped. During each of the glacial Ages, Cooke believed the land stood relatively high and was being eroded. The Pamlico shore line and associated sediments he assigned to the Mid-Wisconsin interval. Cooke's Pamlico shore line is 50 feet lower than the writer's estimate of Mid-Wisconsin sea level. Cooke also recognized a questionable Silver Bluff shore line about 5 feet above sea level and suggested that this also belongs to the Mid-Wisconsin interglacial interval. The writer believes this should be placed in the Thermal Maximum. Cooke emphasized that the higher the shore line the older it is. This progressive decrease in the rise of sea level during successive interglacial intervals Cooke does not explain as resulting from decreasing degrees of deglaciation but as "a sinking of the bed of the ocean in the North Atlantic, the Indo-Pacific region, off the coast of California, or in some other unstable part of the world" (Cooke, 1945, p. 245).

Flint (1940; 1941; 1947, p. 440) recognized only two (or possibly three) raised shore lines

and assigned different ages to them (Fig. 2). His Suffolk shore line is clearly the Pamlico of Cooke, and his Surry is at the same general level as the Wicomico. Flint explained these scarps on a volumetric basis. His argument for

CLASSIFICATION		THEORETICAL SEA-LEVEL VARIATIONS	T E R R A C E S				
			Cooke	Flint	MacNeil	Vernon	Carlson
Post-Wisconsin	Present	0	Submergence to zero				Pamlico (20-30')
	Little Ice Age	-3(?)					
	Thermal Maximum	+30			Silver Bluff (8-10')	Coastal and Submarine Plain	
	Cochrane	-30					
Late Wisconsin	Mankato Cary-Mankato (Two Creeks) Cary	-300	Erosion			Valley erosion and Lake Flint marl	Penholoway (60-70')
Mid-Wisconsin	Tazewell-Cary (Brady)	+75	Silver Bluff(8') Pamlico(25')		Pamlico(25-30') =Suffolk	Pamlico Terrace(25') and formation	
Early Wisconsin	Tazewell Iowan-Tazewell Iowan Farmdale-Iowan Farmdale	-300	Erosion			Valley erosion	Wicomico (90-110')
Sangamon		+150	Talbot(42') Penholoway(70') Wicomico(100')	Suffolk(20-30') =Pamlico	Wicomico(100') =Surry	Wicomico terrace(100') and formation	Sunderland (130-160')
Illinoian		+450	Erosion			Valley erosion	
Yarmouth		+150	Sunderland(170') Coharie(215')	Surry(100') =Wicomico	Okefenokee(150') =Sunderland	Okefenokee deposits(150')	
Kansan		-450	Erosion			Valley erosion	
Aftonian		+150	Brandywine(270') (Citronelle)	Trail Ridge(240-160') (P)	P	Coharie deposits(220')	Coharie (190-210')
Nebraskan		-450	Erosion			Valley erosion	

FIGURE 2.—THEORETICAL SEA-LEVEL VARIATIONS AND MARINE TERRACES OF DIFFERENT WRITERS ON ATLANTIC AND EAST GULF COASTAL PLAINS

the Yarmouth age of the Surry (Wicomico) was that the Yarmouth was the only interglacial age long enough and warm enough to permit the melting of all the land ice and thus to have caused sea level to rise above the 100-foot level (Flint, 1947, p. 440). However, the writer estimates that sea level might well have risen to +150 feet and have become stationary there during any one of the pre-Wisconsin interglacial ages. The Suffolk (Pamlico) terrace was placed in the Sangamon by Flint (1947, p. 440) on the basis that there was slightly less ice than there is today during this interval, and that at all times during the Wisconsin Age, even during the Brady interval, there was more ice than there is now and sea level must have been lower. Evidently Flint does not agree that the Brady interval was so long and warm as the writer thinks it was. Flint and Deevey (1951, p. 288) argued that the Thermal Maximum was the only time since the Sangamon when there could have been less ice than there is now and sea level would have been higher; they concluded that the Pamlico must be Sangamon. The radiocarbon age of the Pamlico deposit (Libbey No. 105) has been determined at more than 20,000 years (Arnold and Libbey, 1951, p. 115)—too old for the Thermal Maximum.

Flint tentatively accepted the Trail Ridge shore line that slopes northward from 240 feet in north-central Florida to 160 feet in southeastern Georgia. Also, tentatively, he assigned this surface to the Aftonian. Since its formation and before Surry (Wicomico) time the land must have been tilted northward. This shore line was evidently diastrophic rather than volumetric, and it might well be Tertiary rather than Pleistocene.

MacNeil (1950) recognized four of Cooke's shore lines and both of Flint's (Fig. 2). MacNeil placed the Yarmouth Okefenokee sea level at 150 feet, the Sangamon Wicomico at 100 feet, the Mid-Wisconsin Pamlico at 25–30 feet, and the Thermal Maximum Silver Bluff at 8–10 feet (MacNeil, 1950, p. 99). He did not rule out the higher and older shore line of Cooke and Flint. The age assignments of the Wicomico and Pamlico agree with Cooke rather than with Flint. However, MacNeil placed the Silver Bluff in the Thermal Maximum rather than in the Brady. This fits the writer's sea-level figures if one assumes no relative uplift of the land since the Thermal Maximum.

The only direct evidence of age of any of these terraces was supplied by Leverett (1934,

p. 32–34) who traced some Illinoian gravel in Pennsylvania into the Wicomico terrace.

Vernon (1942, p. 15–28; 1951, p. 21–27) agreed with Cooke and MacNeil (Fig. 2). He called upon progressive uplift of Florida in adjustment to deposition of sediment in the Gulf and progressively decreasing deglaciation during the Pleistocene to explain four shore lines of decreasing elevation and age. Vernon agreed with Cooke that erosion of elevated lands occurred during the glacial Ages.

Carlson (1950) recognized five scarps in coastal Alabama but made no attempt to assign Pleistocene ages to them (Fig. 2).

There is convincing evidence of two or more raised shore lines on the Atlantic and East Gulf Coastal Plains and all but Carlson agree that they were most likely developed during interglacial Ages when sea level was high. Agreement is far from complete concerning the number of such shore lines, and dates of origin are at best speculative. Even the shore lines of MacNeil and Vernon, one for each of the interglacial intervals including the Thermal Maximum, cannot be explained by increases in sea-water volume alone. The land must have been raised, or sea level lowered diastrophically during the Pleistocene. The alternative is to assume that each deglaciation was smaller than the next preceding one. The limit of rise of sea level by deglaciation alone is about 150 feet above present sea level. The Coharie and Trail Ridge shore lines are higher than this. It seems likely that this limit should have been reached during the Aftonian, Yarmouth, and Sangamon. In any case no definite relation exists between these emergent shore lines and a reasonable estimate of interglacial sea levels.

Texas Coast

What about old shore lines on the Gulf Coast of Texas? This area has been studied by Deussen, Trowbridge, Barton, Price, Doering, Storm, and others, and such shore lines are largely unreported. Instead, there are three coastal plains—the Beaumont, Lissie, and Willis plains, named for underlying stratigraphic formations. For the most part these plains are not noticeably marked off topographically one from another. Price (1934, p. 949) reported a low scarp in a few places separating the Beaumont and Lissie plains. The Hockley scarp northwest of Houston in some places separates the Willis and Lissie plains. However, Barton (1930a, p. 1302–1303) and Rose (1943, p. 1083–1085) said the Hockley scarp is a structural flexure partly buried by younger nonmarine sediments, and Storm (1945, p. 1316) believed it is merely the inner edge of the Lissie delta plain. All three plains are clearly depositional surfaces and the sediments that underlie them are nonmarine. Barton (1930a, p. 1303–1314) considered the Beaumont plain the surface of coalescent deltas only slightly older than the almost non-existent deltas in bays at the mouths of present streams, and by inference considered the Lissie and Willis plains older compound deltas of the same sort. Price (1933b, p. 919–921) described the Ingleside terrace in the Corpus Christi area as an extension of the coalescing delta plain of Barton ending in Bay bluffs 12 feet high. Its age and correlation are uncertain. Doering's (1935, p. 653–654) idea was much the same as Barton's, but he called the plains coalescent fans rather than deltas. In any case they were shaped by streams and not by the sea. Flint (1947, p. 438) suggested that marine shore lines and sediments similar to those on the Atlantic and East Gulf coastal plains were formed in Texas and buried under the sediments that compose the coalescent deltas or fans. But these plains and their subjacent formations are in themselves records of Pleistocene (and possibly late Pliocene) history. They are fluviatile rather than marine.

Fluvial Terraces in Mississippi River Valley

Fisk-Russell Interpretation

Fisk (1938a; 1938b; 1939a; 1939b; 1940; 1944; 1949) and Russell (1940) outlined a history for the alluvial valley of the Mississippi River that accounts for the main glacial and interglacial Stages and Substages of the Pleistocene. The south-flowing Mississippi River Fisk believes came into existence only with the advance of the Nebraskan ice sheet;

he believes its history was wholly Pleistocene (Fig. 3). Four stream terraces (Williana, Bentley, Montgomery, and Prairie) and five cycles of valley cutting are recognized. The Williana, Bentley, and Montgomery terraces

Now		
Late Wisconsin	Post-maximum	Alluvial Valley Filling
	Maximum	Valley Cutting 5
	Pre-maximum	
Mid-Wisconsin		Prairie Alluviation
Early Wisconsin	Post-maximum	
	Maximum	Valley Cutting 4
	Pre-maximum	
Sangamon		Montgomery Alluviation
Illinoian	Post-maximum	
	Maximum	Valley Cutting 3
	Pre-maximum	
Yarmouth		Bentley Alluviation
Kansan	Post-maximum	
	Maximum	Valley Cutting 2
	Pre-maximum	
Aftonian		Williana Alluviation
Nebraskan	Post-maximum	
	Maximum	Valley Cutting 1
	Pre-maximum	

FIGURE 3.—HISTORY OF ALLUVIAL VALLEY OF MISSISSIPPI RIVER ACCORDING TO FISK AND RUSSELL

are benches sloping gently downstream, composed of chert gravel overlying bedrock formations of different ages. Such gravels were considered the inland and upstream representatives of the Pliocene Lafayette formation by McGee (1891) and more recently were known as Citronelle (Matson, 1916, p. 168, 172). The Citronelle was dated by a Pliocene flora from clay at Lamberts Station 5 miles south of Citronelle, Alabama (Berry, 1916, p. 195). Roy (1939, p. 1557) reported that the clay containing these fossil plants was faulted up against the gravel and is older than the gravel; he advocated abandonment of the term Citronelle (Roy, 1939, p. 1559).

Fisk recognizes no post-Miocene and pre-Pleistocene sediments in the lower Mississippi Valley and considers the gravels associated with the Williana, Bentley, and Montgomery terraces Aftonian, Yarmouth, and Sangamon respectively. The Natchez formation (Chamberlin, 1896) in and near the type locality consists of silt, sand, and some gravel and is locally as much as 100 feet thick. The sand and gravel contain glacially derived materials, and the formation is doubtless Pleistocene. The Natchez overlies gravel that resembles Citronelle. Fisk (1944, Fig. 26) first considered the Natchez part of the Sangamon Montgomery terrace materials, but later (1952, Fig. 3) as a part of the Yarmouth Bentley terrace.

In the Williana and Bentley terraces loess overlies the gravel. According to Russell (1944) and Fisk (1944; 1951) the loess in each terrace is colluvium resulting from the reworking of the back-swamp deposits of each terrace during the next valley cutting. Fisk concludes that as any one of the ice sheets began to form and spread, sea level began to drop and valley cutting started. This erosion continued through the maximum of glaciation. With the retreat of the ice, sea level rose, and the river and its tributaries deposited.

This sequence of events cannot be explained entirely by glaciation and low sea level followed by deglaciation and high sea level, even if it is assumed that during each succeeding interglacial age less ice was melted than during the preceding one. In fact, according to Fisk (1944, Figs. 75, 77), deglaciation took place to the same extent during all the interglacial intervals, including the Late Wisconsin post-maximum of age zero and sea-level zero. This leads to the question of what sort of control there was to leave the same amount of ice as there is today and sea level stabilized at zero at the end of each previous glacial Age. Progressive net uplift of the land relative to sea level is indicated.

Horberg-Leighton-Willman Interpretation

Horberg (1945; 1946; 1950a; 1950b; 1953) and Leighton and Willman (1949; 1950) do not agree with Fisk in several important particulars. In Illinois old erosion surfaces and valley terraces occur (Fig. 4). Lancaster, Central Illinois, and Havana surfaces are found over considerable areas in Illinois under glacial drift. Deep erosional valleys below the Havana level are considered pre-Nebraskan, because sand and gravel believed to be older than Kansan occur in them. Also the three erosional

surfaces are associated with Lafayette-type gravels that contain no foreign glacial material and that are generally considered Tertiary.

The Early and Late Mankato terraces of Illinois are depositional, contain abundant glacial material, and converge vertically downstream. In the Illinois River Valley the two terraces are 15–20 feet apart in elevation; Leighton and Willman (1949, p. 15, 18, 22) believe an early Mankato valley train was deposited, was eroded by the Kankakee Torrent, then a late Mankato valley train was deposited, and more recently erosion changed the last valley train depositional surface into a low terrace.

Pleistocene	Late Mankato Terrace
	Kankakee Torrent
	Early Mankato Terrace
Pre-Pleistocene	Uplift and cutting of deepest valleys
	Havana Strath
	Uplift and erosion
	Central Illinois peneplane
	Uplift and erosion
	Lancaster peneplane

FIGURE 4.—HISTORY OF MISSISSIPPI VALLEY AREA ACCORDING TO HORBERG, LEIGHTON, AND WILLMAN

In Nebraska (Schultz and Stout, 1945) and Kansas (Frye and Leonard, 1952, p. 116–140) three loesses of different ages occur (Fig. 5). In Illinois pro-Illinoian Loveland loess, pro-Wisconsin Farmdale loess, and Peorian loess (including loess of Iowan and Tazewell age) are recognized. Bignell loess has not yet been reported in Illinois. The silts of the loesses have been removed by wind, chiefly from glacial depositional surfaces such as valley trains, and have been transported by wind and deposited on the lee sides of the source surfaces. There is abundant evidence for these conclusions.

Leighton and Willman (1949; 1950) traced the surfaces of Horberg and the Mankato terraces from Illinois down the Mississippi Valley into the lower alluvial valley described by Fisk and found that they tie up with Fisk's terraces (Fig. 6). But the Williana, Bentley, and Montgomery are considered by Fisk and Russell Late Nebraskan and Aftonian, Late Kansan and Yarmouth, and Late Illinoian and Sangamon respectively, whereas Leighton and Willman conclude that they are pre-Nebraskan and have nothing to do with glaciation and deglaciation. Leighton and Willman (1949, p. 71; 1950, p. 614) agree with Chamberlin (1896), Chamberlin and Salisbury (1907, p. 386–388), and Vestal (1942, p. 50–59) that the Natchez formation is unconformable on and much younger than the underlying gravel, and that the Natchez formation is a remnant of some pre-Wisconsin valley train.

Late Wisconsin	Bignell loess
Mid-Wisconsin	Brady soil
Early Wisconsin	Peoria loess
Sangamon	Soil profile
Illinoian	Loveland loess

FIGURE 5.—LOESSES AND SOIL PROFILES OF NEBRASKA AND KANSAS

Three loesses (in addition to the Farmdale loess) separated by soil profiles occur on at least the two older terraces in the lower valley. Whether these loesses are the pre-Loveland, Loveland, and Peorian of Leighton and Willman (1949, p. 71; 1950, p. 614–619) or the Loveland, Peorian, and Bignell of Nebraska and Kansas remains to be determined.

Discussion

Leighton and Willman on one side and Fisk and Russell on the other side agree that the Mankato (Prairie) terrace is the youngest of the four and is made of glacial material. However, Leighton and Willman place the age of this terrace as Mankato, and Fisk considers it late Early Wisconsin and Mid-Wisconsin. Leighton and Willman consider the deep erosional bedrock valley pre-Pleistocene. They think the fill in this valley was deposited during the several glacial Ages of the Pleistocene, chiefly by meltwater. The material beneath the

present flood plain they consider part of the Pleistocene valley deposits that has not been removed during successive interglacial erosion intervals (Leighton and Willman, 1949, p. 5). As the bottoms of these erosional valleys are

Fisk and Russell	Horberg, Leighton and Willman
Prairie	Late Mankato Early Mankato
Montgomery	Havana
Bentley	Central Illinois
Williana	Lancaster

FIGURE 6.—RELATION OF MISSISSIPPI VALLEY TERRACES OF FISK AND RUSSELL TO EROSIONAL SURFACES AND TERRACES OF HORBERG, LEIGHTON, AND WILLMAN

at all latitudes far below the levels of the terraces and the existing flood plain, a net subsidence of the land or net rise of sea level is indicated for the Pleistocene. According to Fisk the net Pleistocene movement was in the opposite direction.

The pros and cons of this argument have been stated fully (Fisk, 1938a; 1938b; 1939a; 1940; 1944; 1949; 1951; Horberg, 1945; 1946; 1950a; 1950b; 1953; Leighton and Willman 1949; 1950) and have been summarized above. It should be added that Leighton and Willman (1949, p. 5a) believe the glacial intervals were too short to have allowed the Mississippi River and its tributaries to cut deep and long valleys while sea level was low. They (1949, p. 78–79) list mineralogical studies made by others showing that the Pleistocene tills and loesses in the North-Central States, the Mankato (Prairie) terrace, and the loesses in the lower valley are rich in ferromagnesian minerals. They state that these minerals were brought by the glaciers and glacial meltwater from northern crystalline rocks. In contrast they claim that the gravels associated with the Williana, Bentley, and Montgomery terraces and with the corresponding Lancaster, Central Illinois, and Havana erosion surfaces contain few such minerals but an abundance of kyanite and staurolite similar to known Cretaceous and Tertiary formations. Fisk (1939b) reported 27 igneous and metamorphic rock types, chiefly granites, gneisses, and quartzites, in gravels dredged from deep pits in central Louisiana; he considered them derived from the Lake Superior Upland and the Central Interior Region in Pleistocene time. The rarity or absence of such rocks in the outcropping upper portions of the gravels Fisk attributed to weathering. The gravels studied came from 11 pits in 6 parishes and represent all 4 of Fisk's terraces. Fisk *et al.* (1949, p. 38) states that 10 per cent of the pebbles and cobbles contained in the gravels consist of igneous and metamorphic rocks. Fisk (1951, p. 341–343) also argues that considerable mineralogical variation exists both in the gravels and in the loesses, that the gravels contain some minerals and rocks that appear glacial in origin, that much of the gravel was brought in by tributary streams from nonglaciated areas to be mixed with minor amounts of glacial material in the main valley, and that at best correlation of sediments from different sources by mineral composition involves serious uncertainties.

It seems unusual for old peneplanes and other extensive erosional surfaces to finger out into stream terraces in the downstream direction (Horberg, Leighton and Willman) rather than in the upstream direction.

For Leighton and Willman the known events of the Nebraskan, Aftonian, Kansan, and Yarmouth are not recorded in the lower valley. The Natchez formation of unknown age, the Loveland loess, and the soil profile on the Loveland loess constitute the only definite records of pre-Wisconsin history.

The writer believes that the Mississippi valley loesses are mainly eolian but the ages and origins of the terraces are still in question.

Pleistocene History of Driftless Area

For many years graduate students and the writer have studied the erosional history of the Driftless Area in western Wisconsin, northwestern Illinois, and adjacent portions of northeastern Iowa (Trowbridge, 1921; 1934; 1935; Trowbridge, Williams, Frye, and Swenson, 1941). Topographical and sedimentological data have led to certain current

conclusions concerning this particular area (Fig. 7).

A single old erosion surface, the Lancaster peneplane, bevels the Paleozoic strata and slopes gently southward. It is 300–600 feet higher than the alluvial bottoms of present-day valleys and 625–725 feet higher than the bottoms of bedrock valleys beneath sand and gravel fills. In the Driftless Area no upland gravels and sands occur on this surface, except for scattered pebbles derived from an older deposit. The Windrow formation of Wisconsin and Iowa (Thwaites and Twenhofel, 1921), the deposits near Red Wing, Minnesota, and the Ostrander formation of Fillmore County, Minnesota (Stauffer and Thiel, 1944, p. 332) slope westward and are now considered Cretaceous.

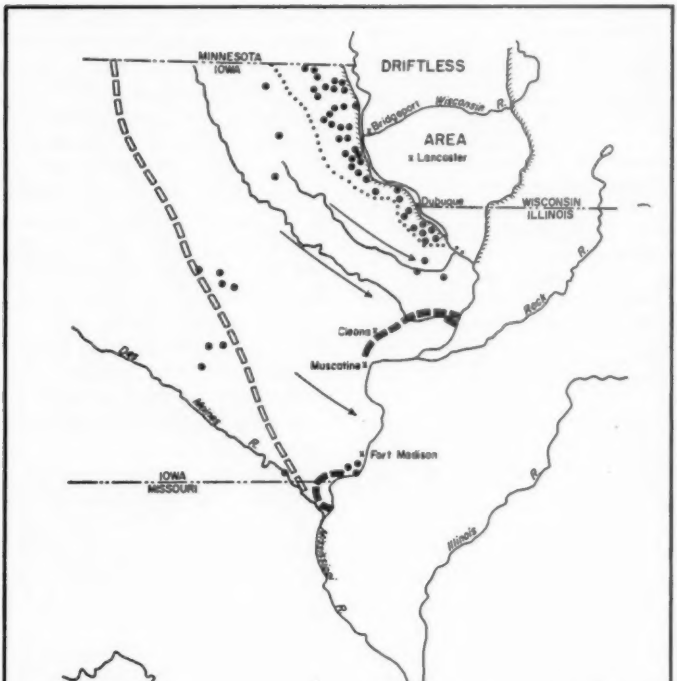

FIGURE 7.—SKETCH MAP SHOWING BORDER OF DRIFTLESS AREA, PRE-NEBRASKAN AND POST-NEBRASKAN COURSES OF MISSISSIPPI RIVER, AND SOME LOCATIONS OF NEBRASKAN DRIFT IN IOWA, ACCORDING TO TROWBRIDGE, WILLIAMS, FRYE, AND SWENSON

In northeastern Iowa remnants of the Lancaster surface are covered discontinuously by patches of Nebraskan glacial till. This till lies everywhere at and above the level of the Lancaster surface or in sinks and caves less than 100 feet below this level. Scores of oc-

currences of this drift are known between the east border of the Kansan drift and the Mississippi River, but none on the east side of the river. The river in this position is an ice-border stream with reference to the Nebraskan drift occurs under Illinoian drift at many places in western Illinois. The Mississippi River was forced from its Aftonian position through a deep valley leading from the neighborhood of Clinton, Iowa, on the Mississippi River to Hennepin, Illinois, on the Illinois River and thence down the Illinois River and lower Mississippi River to the Gulf. That such a drainage pattern existed at some time has been known for many years.

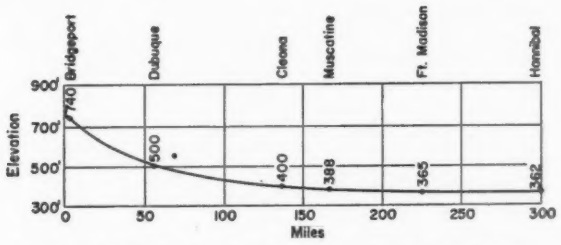

FIGURE 8.—EARLY KANSAN PROFILE OF MISSISSIPPI RIVER
Gradient 1.8 feet per mile Bridgeport to Fort Madison

glacier. Its pre-Nebraskan course is buried farther west under drift of three ages.

No evidence indicates that the glacier that deposited this oldest till affected the valleys. The till is evidently older than the valleys. There is no evidence of the existence of erosional surfaces younger than the Lancaster peneplane, such as the Central Illinois and Havana surfaces described by Horberg in Illinois. However, at Bridgeport and Wauzeka, Wisconsin, on a bedrock bench in the valley of the Wisconsin River, glacial till and fluvio-glacial materials occur. These materials are ascribed by MacClintock (1922, p. 673–680) to the Kansan. This bench seems to represent the depth to which the Mississippi and Wisconsin rivers had cut their valleys by the time the Kansan glacier reached this position. The surface on which the drift lies is 440 feet below the Lancaster surface and 285 feet above the bottom of the bedrock erosional valley adjacent. About 50 miles downstream from Bridgeport, at Dubuque, Iowa, the bedrock bottom of Couler Valley, the abandoned valley of Little Maquoketa River, is 400 feet lower than the Lancaster surface and 224 feet higher than bedrock in the bottom of the Mississippi valley (Trowbridge, 1936). Several points farther downstream tie in with the Bridgeport terrace-Couler Valley profile and indicate the Early Kansan course of the Mississippi River (Fig. 8).

The Kansan drift west of the Driftless Area on the other hand occupies upland flats and valleys alike and was evidently deposited after the valleys were started (Fig. 9). The Kansan drift border crosses Mississippi River at the south end of the Driftless Area, and this

The fill in the Mississippi River valley in the Driftless Area, more than 300 feet thick at Dubuque, is late Wisconsin. The terraces, were traced by Leverett (1921, p. 623–624) far downriver from Mankato marginal moraines near St. Paul, and are known to extend downstream from Cary moraines on Wisconsin River at Prairie du Sac. These also are the Early and Late Mankato terraces of Leighton and Willman and the Prairie terrace of Fisk.

On the whole, the Driftless Area history outlined seems more in harmony with Fisk's idea that the Williana, Bentley, and Montgomery terraces are Pleistocene than with the conclusions of Horberg, Leighton, and Willman that corresponding erosional surfaces are pre-Nebraskan. However, there are important points of difference with both sets of conclusions.

There is evidence that the Mississippi River existed and flowed south at least as early as late Tertiary time and did not come into existence with the advancing Nebraskan ice

sheet. On the other hand, evidence from the Driftless Area places most of the history of the river in the Pleistocene. The conflicting evidence in Illinois that makes the deep-stage erosional valleys pre-Pleistocene, and in Iwoa post-Kansan, has been recognized for some years but has not yet been resolved. The writer believes that, in the upper valley at least, the deposits belong to the glacial Ages and the erosion intervals to the long interglacial intervals. The cutting of valleys 400 feet deep in resistant bedrock 1000 miles from the sea at one time and 250 feet deeper at a later time cannot be attributed to falling sea level that comes with the short time of accumulating land ice and the maximum stand of that ice. The cutting of the valleys in the Driftless Area must have been accomplished during the long interglacial intervals despite the high sea level. The writer believes that Pleistocene regional diastrophic uplifts occurred in the Driftless Area and that they were unrelated to glaciation.

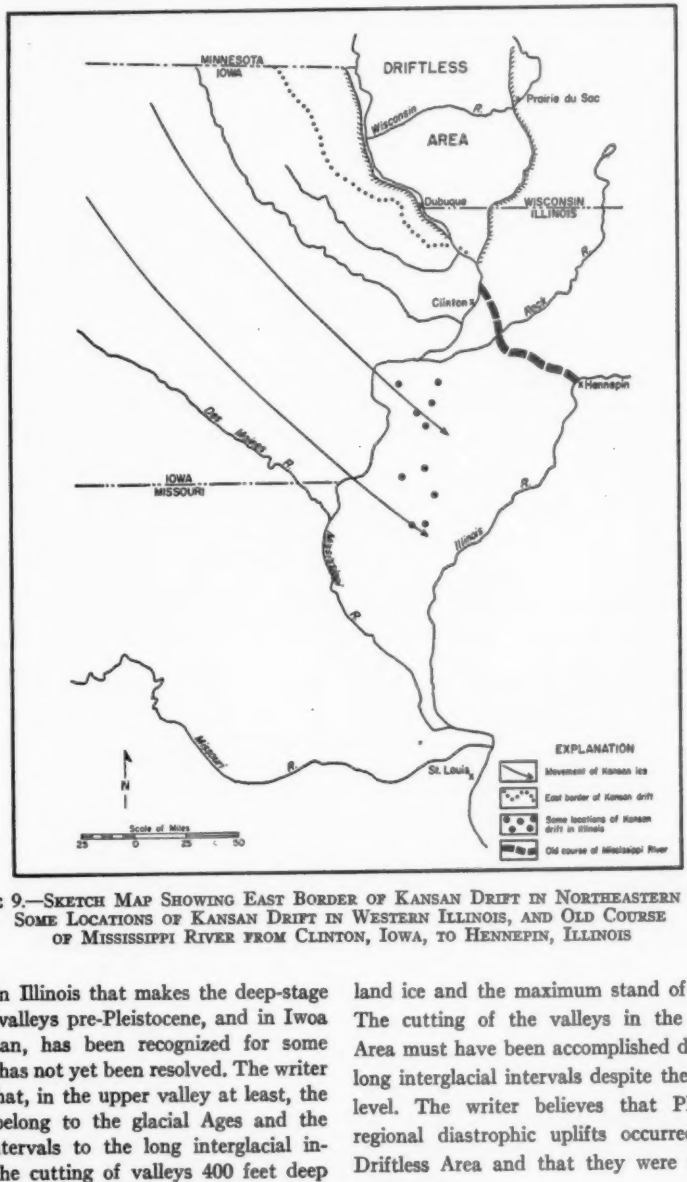

FIGURE 9.—SKETCH MAP SHOWING EAST BORDER OF KANSAN DRIFT IN NORTHEASTERN IOWA, SOME LOCATIONS OF KANSAN DRIFT IN WESTERN ILLINOIS, AND OLD COURSE OF MISSISSIPPI RIVER FROM CLINTON, IOWA, TO HENNEPIN, ILLINOIS

Record of Gulf Coast Sediments

Theoretical Considerations

The most nearly complete record of Pleistocene history should be found in the sediments of the Gulf Coastal area and the offshore Continental Shelf. Provided no other events took place but glaciation and deglaciation and the resulting fall and rise of sea level, the shore lines during the glacial Ages would have been at or near or even beyond the outer edge of the Continental Shelf. At these times, beaches, lagoons, barrier islands, and deltas, would have been formed at and near sea level, and neritic marine deposits would be laid offshore.

The seaward slope of what is now the Continental Shelf was doubtless gentle, as it is now (about 3–5 ft. per mile), and might have been eroded or it may have become the site of deposition of continental sediments. At least this surface would have been emergent and subject to subaerial processes.

However, during the long interglacial Ages, sea level would have stood high on the land, submerging not only the Continental Shelf but also land surfaces now exposed to the air. Along these submergent shore lines, also, normal shore features would have been developed, and offshore the usual shelf sediments would have been deposited. There should have been four of these sets of marine deposits, one for each of the Aftonian, Yarmouth, Sangamon, and Brady, and possibly a fifth set representing the Thermal Maximum.

If erosion took place on emergent surfaces during glacial Ages, four or five marine formations separated by disconformities would have resulted. But, if the exposed surfaces were so low and flat that they received deposits, then the Pleistocene should consist of 9 or 11 alternating marine and nonmarine formations.

How well does what is known of Gulf Coast and Shelf sediments fit this picture?

Mississippi River Delta Area

Fisk and his associates (1952), as results of studies in the Atchafalaya Basin of Louisiana, outline a history of the Mississippi Delta that includes all the main events and conditions of the Pleistocene. Again there is no record of nonmarine Pliocene history, for the equivalents of the Citronelle and the Willis are considered Pleistocene.

The Pleistocene history of the Delta area, according to these writers, checks closely with Fisk's history (1944) of the alluvial valley of the Mississippi River. The history starts in the Nebraskan glacial Age when sea level was low and the valley was deeply intrenched. Then in Aftonian time sea level was high (about at zero), the Williana formation was deposited, and the Williana coastwise deltaic plain and river terrace were formed. Again in Kansan time sea level was low and valley cutting took place. In the Yarmouth interglacial interval the river deposited the Bentley formation and formed the Bentley deltaic plain and terrace. This was followed by Illinoian valley cutting, Sangamon deposition of the Montgomery formation, Early Wisconsin valley cutting, Mid-Wisconsin deposition of the Prairie formation, Late Wisconsin valley cutting, and post-Wisconsin deposition of Recent alluvium.

On the outcrops these formations are fluvial and nonmarine, but down the dip they interfinger with marine deposits. The Prairie formation contains brackish-water and marine megafossils at the surface near the coast, and at several horizons beneath the surface as recorded in wells (Richards, 1939a).

Plate 4 of Fisk's report of 1952, which is an east-west section near the lower end of the Atchafalaya basin and on the Prairie hinge line shows the recent alluvium as low as 450 feet below sea level, the Prairie formation to −675 feet, the Montgomery to −990 feet, and the bottom of the Bentley below −1200 feet. Presumably the Williana formation lies unconformably below the Bentley and at still lower subsea levels. Each formation attributed to an interglacial Age consists of a sandy and gravelly substratum overlain by a sand, silt, and clay topstratum. Fisk (1944, Figs. 69, 70) locates the base of the Pleistocene at −4000 feet under the lower part of the Mississippi Delta complex. He shows (Fisk, *et al.* 1952, Fig. 20, p. 75) the Recent topstratum deposits as 50 feet thick at Donaldsonville, 500 feet thick at Venice, and 700 feet thick at Port

Eads. Lowman (1949, Fig. 32, p. 1990) showed the base of the Upper Marine Beds (Pleistocene) at depths of 1500 feet at St. James, Louisiana; 2500 feet in central Terrebone Parish; 3000 feet near the head of Timberlier Bay; and 3780 feet at Timberlier Pass. This horizon Lowman considered close to the base of the Pleistocene. It would be interesting to know whether the Foraminifera in these beds occur in either alternating cold- and warm-water zones or alternating shallow- and deep-water zones.

A large amount of net compaction or subsidence must have occurred during the Pleistocene on the Gulfward side of the hinge line and a net uplift of less magnitude on the landward side. The Pleistocene subsidence near the present coast represents a continuation of the Gulf Coast geosyncline.

In this report on the Atchafalaya Basin area (Fisk *et al.*, 1952) special attention is given to Late Wisconsin and Post-Mankato history. During Late Wisconsin glaciation the shore line was near, at, or beyond the head of the offshore submarine canyon west of the present delta, and deep trenches were cut in the Coastal Plain formations by the main stream and in more resistant rocks by tributary streams. Coarse materials resulting from this erosion were swept into the sea (Fisk *et al.*, 1952, p. 35–37). With the melting of the Wisconsin ice and the resulting rise of sea level, the stream gradients were decreased, and the tributaries overloaded the main stream which deposited and became braided. About 5000 years ago the Wisconsin ice melted, a condition of standing sea level was reached at about the present level, and the river changed from a braided to a meandering stream and became poised (Fisk *et al.*, 1952, p. 39–48). Within the last 5000 years, with no further change in sea level, the Mississippi River changed its course several times, taking shorter courses to the sea, and deposited several successive subdeltas.

These conclusions, based on a huge mass of basic data, are convincing and are generally accepted on the Gulf Coast. The valley-cutting intervals are very short, being confined to the waxing of the glaciers, and the principal valley filling is also short and confined to the waning of the ice sheets. The long interglacial Ages of standing sea level and poised streams are hard to explain unless all the land ice of the world melted, in which case interglacial sea level would have been about 150 feet higher than present sea level rather than at present sea level. There is again no record of fluctuating climate since the Late Wisconsin glacial maximum. According to this interpretation the Mankato ice gradually melted, sea level gradually rose, the melting ceased, and sea level became stationary at zero 5000 years ago, 1000 years or so before the end of the Post-Mankato Thermal Maximum. Within this 5000 years many or most of the present glaciers were formed, the Greenland and Antarctic ice sheets were enlarged, and sea level must have dropped; today all these ice masses are shrinking, and sea level must be rising.

It should be emphasized again that this history involves more than glaciation and deglaciation and the resulting fall and rise of sea level. The coastal areas subsided greatly during the Pleistocene, and the inland surfaces were uplifted and tilted slightly.

Florida

The Pleistocene sediments of Florida that are associated with old shore lines now above sea level, as described and interpreted by Cooke, Flint, Vernon, and MacNeil, have already been discussed. Five additional Pleistocene formations are widely distributed in southern Florida.

The Anastasia formation, the Miami oölite, and the Key Largo limestone are marine, at least in part contemporaneous, and are probably chiefly Sangamon (Cooke and Mossom, 1929, p. 199–210; Parker and Cooke, 1944, p. 66–71; Cooke, 1945, p. 256–272). The Key Largo is a discontinuous reef in the Florida Keys, 90 miles long from Miami to Bahia Honda and 1–3 miles wide. Including parts above and below sea level the reef may be 200 miles long (Cooke and Mossom, 1929, p. 210). The limestone is made up of corals (about 25%), calcareous algae, mollusks, Foraminifera, and echinoids. The Anastasia formation and Miami oölite were deposited in shallow water between the Key Largo reef

and the Sangamon shore line, which was 150 miles farther north (Cooke, 1945, p. 263).

The Fort Thompson formation records a large part of Pleistocene history. It consists of alternating beds of marine- and fresh-water sediments and has been described and interpreted by Cooke and Mossom (1929, p. 211–215), Parker and Cooke (1944, p. 72–74, 89, 94–96), and Cooke (1945, p. 249–256). The Pliocene-Pleistocene contact, the Aftonian, Kansan, Yarmouth, Illinoian, and Sangamon are all represented. The interglacials are marine, and the glacials are recorded by fresh-water marls and unconformities.

The Lake Flirt marl (Cooke, 1945, p. 311–312) was deposited in fresh-water lakes. It overlies the Pamlico and older formations and must be very young; indeed it may still be accumulating.

The Citronelle of Florida is placed in the Pliocene (Cooke, 1945, p. 229–238). Fresh-water diatoms from Florida peats are said by G. Dallas Hanna (1933, p. 74–75) to have grown under climatic conditions now existing in the northern United States and Canada, indicating a glacial climate.

The total thickness of the Pleistocene of Florida probably does not exceed 150 feet (Toulmin, 1952, Pl. 1; Fig. 8).

Alabama

In Alabama the Citronelle formation is listed by Carlson (1950, p. 1119–1120) as either late Pliocene or early Pleistocene. It is 40–100 feet thick near the coast (300 feet at Citronelle) and dips 6–8 feet per mile toward the Gulf; the next younger sediments of the Coharie formation are horizontal. After the deposition of the Citronelle and before the Coharie was deposited, the Citronelle was elevated, tilted, and eroded. The pebbles in the Citronelle consist of quartz, quartzite, and chert. There being no other adequate source of chert (known to Carlson), these pebbles are thought to have been contributed by the Mississippi River (Carlson, 1950, p. 1120). It seems, however, that there is abundant chert in Ordovician, Mississippian, and Cretaceous dolomite, limestone, and chalk to the north, from which the Citronelle chert could have been derived, without raising the question of the age of the Mississippi River.

Mississippi

In Mississippi, in addition to High and Low depositional terraces, the upper part of the Graham Ferry formation is tentatively placed in the Pleistocene (Brown *et al.*, 1944, p. 45–59), and the overlying Citronelle is called Pleistocene without question. The upper part of the Graham Ferry contains abundant ferromagnesian minerals similar to those brought by the Pleistocene Mississippi River to the Gulf Coastal Area (R. J. Russell, 1936, p. 380–381; R. D. Russell, 1937; Goldstein, 1942, p. 81–82). If these minerals are of glacial origin and were contributed by the Mississippi River, then the Graham Ferry, or the upper part of it, is really Pleistocene, and of course the Citronelle is Pleistocene also. This is in accord with the interpretations of Fisk. Possibly, however, the ferromagnesian minerals were derived from crystalline rocks of the Central Mineral Region of Texas (Cogen, 1940, p. 2101).

However, Fisk (1951, Fig. 2) maps the Williana, Bentley, Montgomery, and Prairie formations and terraces in Mississippi as well as in Louisiana, and concludes that the two states have had identical Pleistocene histories.

Texas

Materials now included in the Willis, Lissie, and Beaumont formations of the Texas coastal area were described and interpreted by Penrose (1890, p. 59–64), Dumble (1894, p. 560–566), Deussen (1914, p. 77–83; 1924, p. 102–113), Trowbridge (1926; 1932, p. 184–218), Plummer (1932, p. 749–795), Weeks (1933), Price (1933b) and others. More recently these same formations have been discussed by Meyer (1939, p. 187–192), Metcalf (1940), Cogen (1940), Rose (1943, p. 1085–1089), and Storm (1945, p. 1314–1318). The Willis has generally been considered Pliocene but may be early Pleistocene. The Lissie and Beaumont are Pleistocene. On the outcrops all three are non-marine and fluviatile. The Willis consists of lenticular sands, gravels, silts, clays, and fresh-

water limestones and lies nonconformably above the beveled edges of older formations. The Lissie, also sandy and gravelly, is the result of the erosion and reworking of the Willis. The Beaumont is primarily clay but also contains lenses of sand. It lies nonconformably and disconformably on the Lissie. All agree that these three formations resulted from movements in which uplifts occurred on the landward sides of faults or flexures such as the Balcones and Hockley scarps and relative subsidence, without submergence, nearer the coast. The elevated surfaces were eroded, and the relatively depressed areas received deposits. The shallow pre-movement valleys on the lowlands were filled, and the deposits spread over interstream areas, making continuous deposits of coalescent fans and flood plains. Closer to the coast and downdip from the outcrops, drillings indicate that the formations thicken greatly and grade into or interfinger with marine deposits. Brackish-water and shallow-water marine shells, chiefly *Rangia cuneata* and *Ostrea virginica*, similar to forms living today in coastal waters are found in several exposures and numerous drillings in coastal Texas counties (Richards, 1939b). Some occur at the surface, and others to depths of at least 940 feet. Advances and retreats of the sea, and net subsidence are recorded.

Therefore the Pleistocene formations on the Texas coast can be explained without calling on repeated variations in sea level from far below to far above present sea level. Either this coastal area has not had such a history, or there is no evidence of it. Pleistocene history appears to have been essentially a continuation of Tertiary history. The Willis and Lissie depositional cycles were interrupted by uplifts of the land and tilting that permitted erosion of previously deposited sediments. Such breaks in deposition are common in the Tertiary. The Willis-Lissie break may be pre-Pleistocene. At best there is evidence of only three rather than five or six erosional breaks. The Gulf Coast geosyncline dating back at least to the early Tertiary continued to subside during the Pleistocene.

In the lower portions of all of the stream valleys on the Texas coast deep alluvial fills indicate that in relatively recent time uplift of land or sinking sea level and erosion occurred, followed by deposition that suggests sinking land or rising sea level. This may record Late Wisconsin glaciation and Post-Mankato deglaciation or possibly simply another diastrophic break.

Studies of heavy minerals in Tertiary and Pleistocene sediments of the Texas and Louisiana coastal area (Cogen, 1940, p. 2044–2101) revealed the presence of a zone composed of hornblende and other ferromagnesian minerals in the outcropping Fleming, Willis, Lissie, and Beaumont formations in the basin of Colorado River. The minerals of this zone are similar to those found in the subsurface formations of Louisiana, the Graham Ferry formation of Mississippi, and to those in the several modern subdeltas of the Mississippi River (R. D. Russell, 1937). But those in the Colorado River basin occur in both Tertiary and Pleistocene formations and evidently did not come down the Mississippi but down the Colorado from the Central Mineral Region of Texas. From this source the sediments may have been spread by streams and shore currents to Louisiana and even to Mississippi.

Again, however, Fisk and his associates (1949, Table 2) have divided the Beaumont plain and formation into two plains and formations. The older is correlated with the Montgomery, and the younger with the Prairie of Louisiana and Mississippi. The topographically lower and stratigraphically higher and younger unit like the Prairie "... consists of a marine plain with a thin veneer of river deposits" (Fisk, Personal communication, February 15, 1954). On this basis, the Willis is Williana, the Lissie is Bentley, the Lower Beaumont is Montgomery, and the Upper Beaumont is Prairie. By this correlation the five-fold division (including the Recent alluvium) of the Pleistocene is carried beyond the Mississippi Delta area and the influence of Mississippi River. And still the history has been the same. The coastal areas have been tilted by subsidence, and the landward portions have been elevated and eroded. This raises a question of the causes of the Gulf Coast geosyncline. It subsided in Mississippi and Louisiana during the Tertiary when the Mississippi River did not exist (according to Fisk)

and also in Texas far from the mouths of the River during both Tertiary and Pleistocene.

Sand and Gravel in Deep Water Offshore

It has long been known that at or near the outer edge of the Continental Shelf in the Gulf of Mexico and elsewhere in the world coarse sands and even gravels occur. Collections made by the U. S. Fish Commission steamer ALBATROSS in 1883–1888 in the northern part of the Gulf of Mexico include such deposits. For instance we find (Canu and Bassler, 1928, p. 9): "D. 2387. Gulf of Mexico, northern part; 29° 24′ 00″ N.; 88° 04′ 00″ W.; 32 fathoms; sand, gravel, broken shells; March 4, 1885". Several other samples collected at about the same time and place are described as coarse and fine sand and broken shells.

From Gulf bottom samples offshore from Southwest Pass of the Mississippi River, collected by the U. S. Coast and Geodetic Survey in 1921, it was determined (Trowbridge, 1930, p. 891–892) that 20–25 miles offshore and extending for at least 20 miles roughly parallel to the bottom contours, in 60–110 fathoms of water, sand exists that is coarser than anything now being brought to the Gulf by the Mississippi River. Within this belt a mound or dome occurs, standing more than 100 feet above its surroundings. This dome was later mapped in detail by the Coast and Geodetic Survey and found to be 4 by 5 miles in diameter and 150 feet high, with its base in 350 feet of water (Shepard, 1937, p. 1355). Shepard (1937, p. 1357) also mentioned at least 26 domes farther west in a belt 180 miles along the inner edge of the Continental Slope. Carsey (1950, p. 376) reported that 164 of these domes are known. They may be salt domes. Coral and other organisms occur on the tops of many of them.

As these coarse sediments are separated from shore by much finer sediments, and there is no known source for them on the lands of today, it is generally concluded that older shore lines existed about where the coarse materials now are and that the strand line or continental deposits have been submerged more recently by a rise in sea level amounting to several hundred feet. This explanation harmonizes with the idea of a glacial Age during which sea level was low followed by an interglacial or post-glacial Age of high sea level. This would mean that the older shallow-water deposits have not been buried since their submergence. Presumably the coarse sediments are Late Wisconsin, and the rise in sea level is post-Mankato. Probably no such lowering of sea level could have resulted from the readvance of Cochrane or Little Ice Age glaciers.

Well-rounded sand grains in patchy shelf sediments in the northwestern Gulf of Mexico have been explained tentatively (Stetson, 1953, p. 22–23) as old dune sands that went through several cycles during Pleistocene low-water stages, with the dunes flattened out during later rise of sea level.

Bioherms in Deep Water Offshore

Several investigators have mentioned organic reefs in deep water far from shore in the Gulf of Mexico (Shepard, 1937, p. 1356–1357; Jordan, 1952, p. 741; Stetson, 1953, p. 20–22). The fact that some of these reefs occur in deeper water than that in which the reef-building organisms are known to flourish suggests again a time of lowered sea level and then a rise of sea level. These too may record Late Wisconsin glacial waxing and waning.

Faunal Zones

Direct evidence of minor climatic fluctuations was presented by Phleger (1951, p. 73–81), who studied the Foraminifera in short cores collected in the western Gulf of Mexico in 1947 by the Woods Hole Oceanographic Institution research vessel ATLANTIS. In several of these cores Phleger found Foraminifera in alternating colder- and warmer-water zones. Two or three cold-water zones and one or two warm-water zones are recorded in cores less than 10 feet long. According to Phleger (1951, p. 79): "These are presumed to represent substages and intersubstages of the last glacial stage." Some such interpretation seems plausible. The alternation of colder- and warmer-water faunas might well record advances and retreats and readvances during the waning of the Mankato ice sheet. For

instance one cold fauna might represent the Mankato maximum, the next younger one the Cochrane readvance or halt, and the third one the Little Ice Age. Although the present day is close in time and climate to the Little Ice Age climax, the climate now should be somewhat warmer than it was at either the Cochrane or the Mankato maximum. Neither of the pre-modern warm-water faunas indicates a climate much if any warmer than the modern climate, and therefore no record is afforded of the Thermal Maximum. It thus appears that even these Foraminfera do not fit the recorded Mankato and Post-Mankato history.

It might well be that these alternating colder and warmer faunal zones are much younger than suggested above and are due to minor cyclic or even possibly seasonal variations in modern climate. They might also be due to modern changes in the directions of Gulf currents. Such an interpretation is suggested by the extreme thinness of the sedimentary section containing these zones. Elsewhere in the Gulf coastal area several hundred feet of sediment has been deposited since the climax of Mankato glaciation (Mississippi River Delta Area section).

Conclusions

The general conclusion is that Wisconsin history of north-central United States is recorded to some extent by stream and marine terraces and sediments farther south, but that evidence for pre-Wisconsin glaciation and deglaciation is lacking or uncertain. Satisfactory evidence of Post-Mankato climatic changes (other than simple progressive melting) is sparse. Convincing evidence exists that all parts of the subject area either subsided or were elevated tectonically during the Pleistocene.

It is hoped that in the near future complete cores from the present surface of deposition to the base of the Pleistocene will be available from drillings on land or in the tidelands of the Gulf Coast. The interpretation of such cores might lead to a more nearly complete history of the Pleistocene that will harmonize with and confirm the history as determined from glacial deposits and interglacial weathering products in the glaciated areas.

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Short Notes

SEDIMENTARY FEATURES OF MARTINSBURG SLATE, NORTHWESTERN NEW JERSEY

BY FRANKLYN B. VAN HOUTEN

INTRODUCTION

In northwestern New Jersey the lower part of the Ordovician Martinsburg slate[1] consists chiefly of alternating layers of black shale and gray sandstone characterized by a distinctive suite of small-scale sedimentary features, particularly cross-stratification, convolute bedding, and graded bedding. Flow markings and load casts, intraformational fragments, and sandstone dikes are also present.

This assemblage is remarkably like small-scale features found in well-known marine deposits of Ordovician age in southern Scotland (Henderson, 1935), of Silurian age in Wales (Rich, 1950), of late Mississippian and early Pennsylvanian age in the Ouachita Mountains (Bokman, 1953), of Mesozoic and Cenozoic age in Alpine geosynclinal basins (Beets, 1946; Kuenen and Carozzi, 1953), and of late Cenozoic age in the Ventura Basin of southern California (Natland and Kuenen, 1951; Kuenen, 1953). In contrast, however, Martinsburg deposits contain no large-scale intraformational deformation, no detrital conglomerate, and only rare intraformational fragments. These frequently associated features are generally believed to be products of turbidity currents. Inasmuch as the Martinsburg assemblage may also be ascribed to ancient turbidity currents, the features are described here briefly.

The illustrated examples were taken from a road cut along New Jersey State Highway 94, about 2 miles south of Columbia and 5 miles south of the Delaware Water Gap. In this discussion the terminology for stratification is that proposed by McKee and Weir (1953).

[1] In order to emphasize the fact that all the features described in this paper are of sedimentary origin and are not products of metamorphism, the fine-grained rock is called *shale*.

DESCRIPTION

Throughout the section, gray sandstone and black shale are segregated into distinct and persistent layers from a fraction of an inch to a few feet thick (Pl. 1, fig. 1). The sandstone layers commonly contain subordinate laminae of shale, however, and many thin layers, which appear parallel and persistent in general view, are actually quite irregular (Pl. 1, fig. 5), resembling delicately irregular laminae in Pliocene deposits of southern California (Kuenen, 1953, Fig. 3). As in many of the formations with similar sedimentary features, the alternation of shale and sandstone, referred to as "rhythmic", is a very conspicuous large-scale feature (see Rich, 1950, Fig. 2).

All the small-scale features in Martinsburg rocks, except sandstone dikes and some intraformational fragments, occur in poorly sorted, slightly calcareous sandstone that ranges in grade size from medium sand to coarse silt.

Graded bedding is rather common in the sandstone layers, but it is not a prominent feature because no great range in grain size is involved in the vertical sorting (Pl. 1, fig. 3). The base of the graded strata generally is an uneven surface on which flow markings are preserved (Pl. 1, fig. 2). These structures are protrusions as much as half an inch deep and 1½ inches long, filled with the coarsest-grained sand in the graded beds. Moreover, the markings commonly are asymmetrical in long profile and are oriented in a general northward trend. Many of the thicker graded layers with flow-marked bases show but little intraformational deformation. Similar flow markings are abundant on the bottom of siltstone layers in Silurian deposits of Wales (Rich, 1950, Fig. 4, p. 722–729); they are also present on the basal surface of sandstone and siltstone in the

Stanley and Jackfork formations (Bokman, 1953, Pl. 1C), and of sandstone of the Oligocene Flysch of Switzerland (Kuenen and Carozzi, 1953, Pl. 2B). These markings are believed to be casts of grooves cut in mud by turbidity currents. Other, less regular, small-scale pockets of sandstone in the underlying shale (Pl. 1, fig. 3) resemble load casts, which Kuenen (1953, p. 1058) attributes to plastic deformation of mud under a load of sand deposited by turbidity currents.

In contrast to the irregular flow-marked base of the sandstone layers, the upper surface is more regular and commonly is conformably overlain by shale.

The only protrusion of sandstone into the overlying shale is in the form of sandstone dikes. These are slender, irregular structures, some of which trace down to disturbed areas in the underlying sandstone (Pl. 1, fig. 6).

Most of the Martinsburg sandstone is distinctly laminated, owing to the presence of very thin layers of black clay. Some of the stratification is a parallel or horizontal type. Much more commonly, however, it is cross-stratification composed of very small-scale cross-laminae in sets less than half an inch thick. Some sets are tabular and planar, or current bedding, with the cross-laminae dipping northward (Pl. 1, fig. 4). Most sets are a lenticular trough type with cross-laminae of varying thickness and with no constant direction or amount of dip (Pl. 1, fig. 6). This "scour and fill" or current-ripple type of cross-stratification is also common in the Ordovician deposits of southern Scotland (Henderson, 1935, p. 497), in Cambrian rocks of Wales (Kuenen, 1953, Fig. 2), and in Pliocene rocks of southern California (Kuenen and Carozzi, 1953, Pl. 2A). The presence of tabular planar sets and lenticular trough sets in the same cosets and compound sets suggests that these two kinds of cross-stratification were formed under rather similar conditions of deposition. Moreover, as Henderson pointed out many years ago, the association of small-scale cross-stratification with graded bedding is more common than Bailey (1930, p. 85) and others have claimed. Significantly, Henderson concluded that this association resulted from "deposition during a decreasing bottom current".

Sedimentary structures produced by intraformational deformation (Shrock, 1948, p.

PLATE 1.—SEDIMENTARY FEATURES, MARTINSBURG SLATE

FIGURE 1.—ALTERNATING LAYERS OF SANDSTONE AND SHALE

Layers of black shale generally are thicker than layers of gray medium- to fine-grained sandstone and siltstone in this section. Dip of strata and of steeper slaty cleavage is southward.

FIGURE 2.—FLOW MARKINGS ON BOTTOM OF LAYER OF GRADED SANDSTONE

Structures are filled with coarsest sand at base of graded bed. Longest marking at right (south) side of specimen is 1½ inches long. Equivalent grooves were present on top of the underlying shale. Length of specimen is 5¼ inches.

FIGURE 3.—LAYER OF GRADED SANDSTONE WITH CONVOLUTE BEDDING

Lower two thirds of layer undisturbed, faintly laminated, and inconspicuously graded from fine to very fine sand size. Coarsest grains are in irregular pocket in underlying shale. Upper third is laminated and contorted. Two remnants of overlying shale preserved on top of specimen. Length of specimen is 2½ inches.

FIGURE 4.—SMALL-SCALE TABULAR PLANAR CROSS-STRATIFICATION

Layers of laminated fine-grained sandstone with coset of tabular planar cross-stratification and sets of horizontal laminae. Unconformable on horizontally stratified shale. Length of specimen is 2¼ inches.

FIGURE 5.—IRREGULARLY STRATIFIED SANDSTONE AND SHALE

Rather persistent layers of crinkled, horizontally laminated black shale interbedded with lenticular layers of cross-stratified and contorted very fine-grained gray sandstone. Disturbance of bedding occurred during deposition. Inconspicuous slaty cleavage dips to left. Length of specimen is 6 inches.

FIGURE 6.—SMALL-SCALE LENTICULAR TROUGH CROSS-STRATIFICATION

Layer of fine-grained gray sandstone with cosets of lenticular trough cross-stratification. Some laminae in middle of specimen were contorted and truncated before deposition of overlying horizontally stratified black shale that contains a very thin, persistent lamina of siltstone. The marked contrast between the shale and sandstone layers is characteristic. The shape of the sandstone dike penetrating the shale apparently has been modified by the deformation which produced the slaty cleavage. Length of specimen is 8¾ inches.

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

FIGURE 5

FIGURE 6

SEDIMENTARY FEATURES, MARTINSBURG SLATE

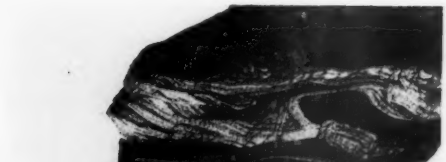

FIGURE 1

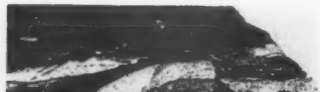

FIGURE 2

FIGURE 3

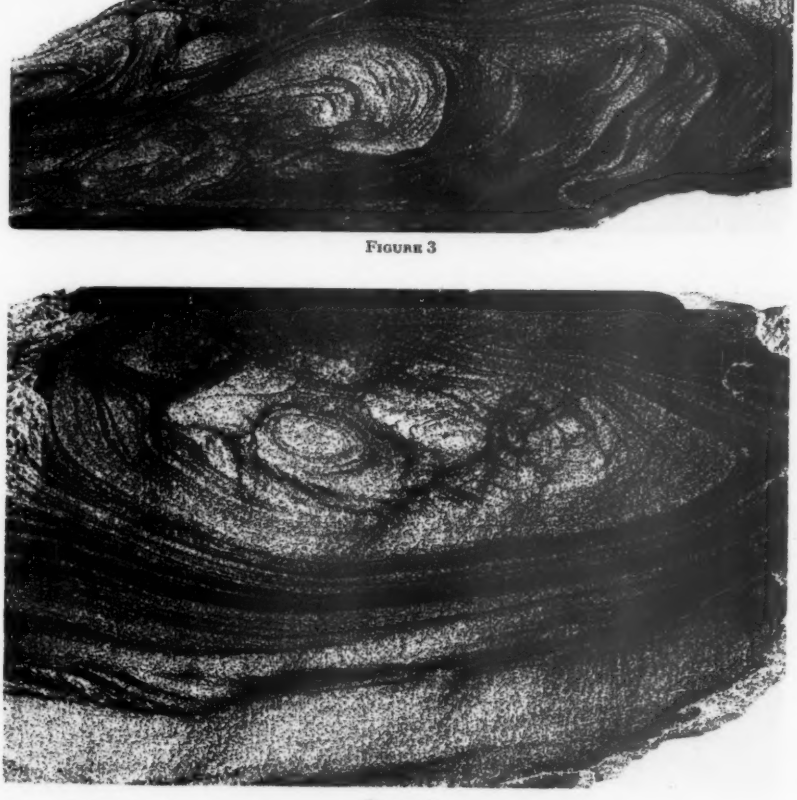

FIGURE 4

SEDIMENTARY FEATURES, MARTINSBURG SLATE

258-275) occur in many of the laminated sandstone layers more than an inch thick. Some less plastically deformed intraformational fragments, or slump structures (Kuenen, 1953, p. 1054–1056), consist of irregular blocks of black shale and siltstone in a matrix of complexly cross-stratified and contorted sandstone (Pl. 2, fig. 1). Others are fragments of limestone and blocks of contorted sandstone in shale and laminated siltstone (Pl. 2, fig. 2). This type of deformation resembles intraformational fragments in Rhaetic beds of Sweden that were produced by subaqueous sliding (Hadding, 1932, Fig. 2, 3). Rarely, angular chips of shale as much as 2 inches long are embedded in contorted and cross-stratified sandstone.

The more common product of intraformational deformation is very intricately contorted bedding and asymmetrical convolutions of stratified sandstone (Pl. 2, figs. 3, 4)[2]. Many of these structures occur in the upper part of sandstone layers that are relatively undisturbed and graded in the lower part (Pl. 1, fig. 3; Pl. 2, fig. 4). Although some layers of contorted sandstone were eroded before deposition of the overlying shale, many are conformably overlain by shale. Similar convolute bedding, characteristically in the upper part of sandstone layers, is very common in Silurian deposits of Wales (Rich, 1950, Figs. 9, 10; Kuenen, 1953, Fig. 9) and also occurs in Ordovician rocks in southern Scotland (Henderson, 1935, Pl. 3, fig. 2) and in Pliocene strata of the Ventura Basin (Kuenen and Carozzi, 1953, Pl. 2A). This kind of sedimentary structure apparently results from deformation during deposition of very mobile sediments (Kuenen, 1953, p. 1056–1057).

In many of the Martinsburg sandstone layers 6 inches to 1½ feet thick, the associated features commonly are arranged in a regular order. A complete sequence consists of the following features:

(1) A surface of erosion cut on the underlying shale
(2) Flow markings on bottom of the succeeding sandstone layer
(3) About an inch of sandstone without lamination; graded and with little clay matrix
(4) Several inches of cosets with parallel laminae of less well-sorted fine-grained sandstone
(5) Several inches of cosets with cross-laminae, more commonly a lenticular trough type
(6) Several inches or more of contorted bedding, ranging from gently deformed to concentric overturned structures
(7) Fraction of an inch of undisturbed laminated very fine-grained sandstone or siltstone
(8) Upper contact generally conformable

[2] Such structures have been called contorted bedding, convolute bedding, intrastratal contortion, and intraformational corrugation.

PLATE 2.—SEDIMENTARY FEATURES, MARTINSBURG SLATE

FIGURE 1.—INTRAFORMATIONAL FRAGMENTS OR SLUMP STRUCTURES

Irregular, elongate blocks of black shale and dark-gray siltstone in a matrix of complexly cross-stratified and contorted very fine-grained sandstone. Delicately scalloped borders of fragments are characteristic. More regular low-angle, lenticular trough cross-stratification in lower part of specimen. Height of specimen is 5¼ inches.

FIGURE 2.—INTRAFORMATIONAL FRAGMENTS OR SLUMP STRUCTURES

Irregular, elongate blocks of gray fine-grained sandstone in a matrix of black shale and dark-gray siltstone. Two smaller, rounder fragments of light-gray limestone (etched) in shale matrix in right half of specimen. Height of specimen is 2⅜ inches.

FIGURE 3.—INTRAFORMATIONAL DEFORMATION AFFECTING ENTIRE LAYER OF SANDSTONE

Convolutions of laminated very fine-grained sandstone. Two thin intraformational chips of black shale lie along axis of fold in lower left of specimen. Length of specimen is 5½ inches.

FIGURE 4.—INTRAFORMATIONAL DEFORMATION AFFECTING UPPER PART OF SANDSTONE LAYER

Convolution in upper part of layer of laminated fine-grained sandstone. Locally cross-stratified and but slightly disturbed in the lower part. Abrupt change from the uppermost undisturbed sandstone to undisturbed black shale is characteristic. Length of specimen is 7⅛ inches.

Conclusion

These sedimentary features suggest that the Martinsburg sandstone layers were deposited by turbidity currents and deformed during accumulation. The turbidity currents believed to have transported the sand must have been small-scale, rather dilute currents of low velocity, in contrast to denser, high-velocity currents that produced abundant intraformational conglomerate, detrital conglomerate, and large-scale slump structures in such deposits as the Ordovician of southern Scotland and the Pliocene of southern California.

In his classic discussion of rhythms in sedimentation Barrell (1917, p. 803-804) explained the alternating layers of black shale and gray sandstone in the Martinsburg formation as the product of storm-wave scour of the bottom of a shallow Ordovician sea, followed by sediment settling during a long quiet period. According to his interpretation, first the mud and then the sand settled to the bottom. The guiding precept of this view, and one that has long dominated geologic thought, is the belief that sand is always a shallow-water sediment, hence

> "marine muds where intercalated with layers of sand indicate that they are shallow-water deposits, since the sand can be transported only in limited depths of water. Where the interstratified sands are thin and regular, this feature indicates spreading by wave action and the sediment was deposited at wave base" (p. 777).

An alternative interpretation of the Martinsburg deposits, presented here for consideration, is based on the belief that sand can be transported intermittently to deeper, quiet-water areas of mud accumulation by turbidity currents moving below wave base. Development of this point of view has depended largely on increased knowledge of the behavior of turbidity currents as an agent of transportation, and on the recognition of an association of sedimentary features that is not readily accounted for by an orthodox interpretation. Bramlette and Bradley (1940, p. 14–16) first suggested that turbidity currents produce graded sediments, and since then many deposits have been ascribed to this agent. Bokman (1953, p. 160) has summarized the newer idea concisely:

> "whereas the graywacke beds are products of turbidity flows, the interbedded shales represent quiet-water marine sedimentation which was proceeding continually. The periodic appearance of coarser material (the graywackes) represents, in a sense, foreign intrusions into this zone of sedimentation."

According to the interpretation proposed here, Martinsburg mud accumulated slowly in a quiet marine environment. Periodically, sand stirred up in the shallow sea was carried to deeper water by small-scale turbidity currents. These sand-laden currents moving into areas of mud accumulation cut flow markings in the muddy sea floor. As the motion of the currents slackened, sediments in the denser lower part of the sheet settled in graded beds, while some of the sediments in the less dense upper part were shifted into complex cross-stratification and convolute bedding. Whether the convolute bedding was truncated before burial or was conformably overlain by mud depended largely on how vigorously the water was agitated by intrusion of the turbidity currents and on how soon a mantle of mud began to accumulate after the turbidity detritus had settled.

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